Structural Geology of North America

HARPER'S GEOSCIENCE SERIES

CAREY CRONEIS, Editor

Structural provinces of North America, shown to the edge of the continental shelf.

Structural Geology of North America

SECOND EDITION

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EDITOR'S INTRODUCTION

A. J. Eardley's *Structural Geology of North America* has, since its publication in 1951, become something of a landmark in the geological literature of the New World. This is demonstrated by the broad base of its foreign sales and the fact that, at home and abroad, the volume has received heavy use by stratigraphers, geophysicists and other specialists, as well as by the structural geologists for whom it was written. Moreover, although originally conceived as a textbook for advanced undergraduates, *Structural Geology* soon became a handy and valued general source book for nonacademic professional and economic geologists.

Dr. Eardley, however, has always considered that his *magnum opus* was somewhat out of date even before the first edition was put through the publishing mill. Accordingly, immediately after the book was issued, he set about the onerous task of revising it. For a full decade now he has devoted a considerable amount of his time and efforts to the current revision. The self-imposed "labor of Hercules" has been particularly frustrating and time consuming because during the fifties numerous basic concepts of structural geology have undergone radical change. Thus, fondly held theories of less than ten years ago are now either discarded or seriously challenged. In addition, a vast quantity of new field data has been accumulating so rapidly that revisions can scarcely keep up with the scientific progress.

Dr. Eardley has taken full cognizance of the rapidly evolving theoretical concepts, as well as of the flood of new information. As a result this edition of *Structural Geology* is far from being a reprint—in many chapters it is so extensively revised as to be essentially a new volume. But in addition, much of the best of the first edition remains, and thus it is likely that this volume will continue to be the standard text and reference work in a subdiscipline of geology that is of prime significance in the proper understanding of all other phases of the subject.

The structural evolution of a continent! Relatively few scientific writers have painted on such a broad canvas as Dr. Eardley. He is something of a rarity even among such artists, for he not only works with a broad brush but also takes pains to fill in the details.

The geological fraternity has been indebted to Dr. Eardley for an excellent compendium on structural geology, and that indebtedness is now increased through an exceptional initial task that has become even better done in its redoing.

CAREY CRONEIS

Rice University
June, 1962

PREFACE
TO THE FIRST EDITION

This book is addressed especially to advanced undergraduates in geology. I doubt that it could have been written on a more elementary level and still presume to use the common terminology of the numerous source publications and the language of the professional geologists. In fact, some instructors may consider the book too advanced for undergraduates. I have endeavored, however, to take such measures as will make it understandable to the student who has had basic courses in mineralogy, lithology, and structural geology. It will be well if he has had a course in stratigraphy in which correlation problems have been discussed and in which some attention has been given to the sedimentary environments and sources.

The reader's attention is lost most frequently by the use of unfamiliar formational, fossil, and geographic names. Generally I have not used formational names in the text but, instead, have referred to the deposits by period, epoch, or stage, and have listed the formational names in charts. This has the advantage of easing the reading of the text and still making the student aware of the many formations in the various parts of the country. At the same time it sets the stage for meaningful stratigraphic studies in other courses.

I have discussed stratigraphic correlations only where necessary, and have relied on the latest authoritative correlations in the literature. Geographic names have been treated with care, and I believe all that have been mentioned are on accompanying maps and figures, or on other well-known maps which are referred to as the occasion arises. Where petrographic research has been referred to, I have attempted to discuss it in such terms that the student with a knowledge of the common rock names will understand.

Several professors who teach structural geology have expressed to me the need for a text that treats structural geology from a regional point of view, but I doubt if the present volume is what they want, or that it can be used as a substitute for the standard textbooks on principles. It may be that in those departments where structural geology is taught as a senior course, the book could be used, and principles could be developed col-

laterally. I think, however, that principles will suffer this way. I have the book in mind for an advanced course in regional or structural geology.

I hope also that the book will prove attractive to professional geologists, because some of the maps and ideas about the many fascinating problems of continental growth may be new to them. I also trust that they will not hesitate to set me right about any errors I have made.

Parts of the North American continent are so well known that it did not seem worth-while to do more than describe them briefly and summarize the conclusions that have been so well presented by others. In certain areas, however, I had to marshal the evidence and present it in some detail in order to sustain an original interpretation. For this reason, all parts of the continent may not seem equally treated. I had to bear in mind the professional geologist as a reader when drawing original conclusions.

A series of paleogeologic maps and paleotectonic maps is included in the book. These, I hope, will be referred to repeatedly. They differ decidedly from the familiar paleogeographic map, and for structural studies are much more illuminating. As geologic studies progress, the maps will undoubtedly bear correction, but I have been impressed repeatedly with the adequacy of our knowledge to date in establishing many important relationships.

Where possible I have referred to late summary reports, and have left the reader to go to these, if he wishes all the original references. Where good summary reports are lacking, I have referred to the basic investigations. Our literature bearing on the structural development of the continent is so extensive that I have been continuously beset by the fear that I have missed an important reference, especially for those regions with which I am least familiar.

The research and writing of this book was done at the University of Michigan, where the geologic library is extensive, the departmental facilities are all that were needed, the time to do research work was abundant, and my former associates on the staff were most helpful and congenial. I remain very appreciative of these facilities and opportunities at the University of Michigan.

Miss Dolores Marsik has helped over several years as typist, and Dr. Ruth Bastanchury Boeckerman has assisted in editorial work and has done the final typing. Mr. Derwin Bell assisted in the drafting of many of the figures and plates.

A. J. Eardley

January, 1951

PREFACE TO THE SECOND EDITION

The second edition is an extensively revised version of the first. Seven new chapters have been added, one on the Precambrian orogenic belts and six on the igneous provinces of the western cordillera. Igneous rocks are accorded a more significant place here than in the first edition. Southern Mexico and Central America are treated in a separate chapter as is also the Canadian Arctic. The colored maps of the summary in Chapter 3 have been extensively revised, and several new ones are included.

Better index maps have been added throughout and an attempt has been made to produce an understandable text independent of outside sources of information. However, such maps as the geologic and tectonic maps of the United States and Canada and the several state maps will be indispensable for instructional purposes and should be available to the student or professional geologist reading the book.

The second edition marks a time of major transition in structural geology. In the past geologists have seen evidence in nearly every mountain system of crustal compression, but now a number of authorities postulate earth expansion, differential uplift, and crustal tension. The folds and thrust sheets are being interpreted as gravity slide phenomena from regions of marked uplift. Vertical movements along with distention and wrenching are considered to be the primary aspects of crustal deformation—not horizontal compression.

The writer sees much in favor of the hypothesis of primary vertical movements and has perhaps accorded it greater attention than some will like. However, he has also attempted to present the geology of the several provinces as the authorities have depicted them. Certain sections of the book, therefore, reflect the orthodox concepts of compression, whereas other parts will seem to emphasize primary vertical movements with secondary folding and thrusting. It will take another ten years to resolve the irregularities and to warrant the preparation of a more definitive third edition.

A. J. Eardley

June, 1962

Structural Geology
of
North America

1.

INTRODUCTION

PURPOSE OF BOOK

The purpose of the book is to describe the structural evolution of the North American continent. The chapters concern the formation and constitution of the mountain systems, basins, arches, and volcanic archipelagos; the beveling of the highlands; and the filling of the basins. In short, they treat of the procession of deformational and sedimentary events. Not only does the book seek to chronicle the crustal unrest of the continent, but it also tries to summarize the supporting evidence.

The igneous provinces and their relation to the tectonic provinces are treated. The advances in geophysics in deciphering deep crustal structure are referred to, and the constitution of the crust in several regions is described. Theories of diastrophism that have been proposed for certain structural systems are summarized, and current concepts of mountain building and continental development are presented where appropriate.

METHOD OF PRESENTATION

The structural history of the continent is one both of time and of geographic position. The major scheme of organization of the book could, therefore, follow one or the other. For instance, if organized on a time basis, all the structural events over the whole continent would be reviewed period by period. If on a geographic basis, the structural history of each major province would be followed from the beginning of Paleozoic time to the present. Neither course when rigidly pursued worked out well, but if the chapter headings are scanned, it will be apparent that emphasis in organization has been placed on geographic position.

The necessity of treating a succession of deformational events in a certain province without serious interruption early became plain, and it was decided that the great mountain systems whose histories run through several periods of time must be treated as units. The growth of the continent in its several provinces has been described first during the Paleozoic, and then, in general, the great structural units of the Mesozoic and Cenozoic have been considered. In the résumé of the structural evolution of the continent, Chapter 3, the paleogeologic and paleotectonic maps are presented, and there the development, period by period, is reviewed.

KINDS OF ILLUSTRATIONS

Considerable effort has been made to illustrate every important point developed in the text. Maps, cross sections, and block diagrams are used. Photographs have little value because the structural features described are usually immensely larger than photographs reveal. If the reader desires to know the nature of the topography, other books with a wealth of photographs should be referred to, such as Fenneman's *Physiography of the United States*, Lobeck's *Geomorphology*, Hinds's *Geomorphology*, and Atwood's *Physiography of North America*.

MAPS FOR COLLATERAL USE

The book is not intended to stand entirely alone. The reader or instructor should have the following maps for ready reference, preferably mounted and hanging on the wall at short range.

The Geologic Map of the United States, 1932 edition
The Geologic Map of Canada, 1957 edition
The Geologic Map of North America, 1946 edition
The Tectonic Map of the United States, 1944 edition
Landforms of the United States, 1939. Map by Erwin Raisz
The Tectonic Map of Canada, 1950
The Geologic Map of South America, 1950

These maps will be referred to repeatedly. Although the book contains many illustrations, it does not reproduce the features of the above maps, and if they are not consulted when referred to, the continuity will be interrupted, the evidence not clearly understood, and perhaps the conclusions not appreciated or properly evaluated.

AUTHORITY FOR STRATIGRAPHIC CORRELATIONS

Most field work in structural geology is based on previous paleontologic and stratigraphic work. A report on the structural geology of an area is not considered worth while unless the formations are dated. The principal method of dating is by the fossils present, and therefore, the structural geologist is dependent upon the paleontologist, except in Precambrian terranes. It is conceivable, but not probable, that a sequence of deformational events could be worked out in a local area without reference to fossils or to nearby stratigraphic columns, but to date the events and to relate them to others in widely separated areas is generally impossible without fossils.

A series of articles has appeared in the last few years in the *Bulletins* of the Geological Society of America that summarize the formational correlations throughout North America for each geologic period. They have been prepared by the Committee on Stratigraphy of the National Research Council, and are taken in this book as authority in relating the numerous orogenic episodes throughout the continent. They are as follows:

Chart No.

1. Cambrian formations of North America, Howell *et al., Bull. Geol. Soc. Am.,* vol. 55, pp. 993–1004, 1944.
2. Ordovician formations of North America, W. H. Twenhofel *et al., Bull. Geol. Soc. Am.,* vol. 65, No. 3, 1954.
3. Silurian formations of North America, C. K. Swartz *et al., Bull. Geol. Soc. Am.,* vol. 53, pp. 533–538, 1942.
4. Devonian formations of North America, G. Arthur Cooper *et al., Bull. Geol. Soc. Am.,* vol. 53, pp. 1729–1794, 1942.
5. Mississippian formations of North America, J. Marvin Weller *et al., Bull. Geol. Soc. Am.,* vol. 59, pp. 91–196, 1948.
6. Pennsylvania formations of North America, R. C. Moore *et al., Bull. Geol. Soc. Am.,* vol. 55, pp. 657–706, 1944.
7. Permian formations of North America, A. A. Baker *et al., Bull. Geol. Soc. Am.,* vol. 71, pp. 1763–1801, 1960.
8. Cretaceous formations of the western interior of the United States, *Bull. Geol. Soc. Am.,* vol. 63, pp. 1011–1044, 1952.
9. Cretaceous formations of the Greater Antilles, Central America and Mexico, R. W. Imlay, *Bull. Geol. Soc. Am.,* vol. 55, pp. 1005–1046, 1944.
10. Marine Cenozoic formations of western North America, C. E. Weaver *et al., Bull. Geol. Soc. Am.,* vol. 55, pp. 569–598, 1944.
11. Cenozoic formations of the Atlantic and Gulf Coastal Plain and Caribbean Region, C. Wythe Cooke *et al., Bull. Geol. Soc. Am.,* vol. 54, pp. 1713–1724, 1943.

Additional correlations charts

Thickness and general character of the Cretaceous deposits in the western interior of the United States, Preliminary Map No. 10, J. B. Reeside, Jr., *U.S. Geol. Survey, Oil and Gas Investigations,* 1944.

Nomenclature and correlation of the North American Continental Tertiary, H. E. Wood, 2nd, *et al., Bull. Geol. Soc. Am.,* vol. 52, pp. 1–48, 1941.

Paleotectonic maps of the Jurassic system, *U.S. Geol. Survey, Miscellaneous Geological Investigations,* Map 1–175, 1956.

Paleotectonic maps of the Triassic system, *U.S. Geol. Survey, Miscellaneous Geological Investigations,* Map 1–300, 1959.

EXERCISES

Four types of assignments and exercises are feasible. The first is the reading and reporting of original articles in the literature. It is hoped that

all articles of outstanding importance are referred to in the text. All publications referred to are listed in the bibliographic index. For emphasis on local areas of interest, the instructor can assign additional publications.

The second type of exercise is the detailing of stratigraphic successions in the different basins and mountain systems. This in itself would constitute an extensive course in stratigraphy, but perhaps for local interest, certain stratigraphic details can be fitted into the structural picture.

The third type of exercise is the assembling from the book of all the structural events that occurred nation-wide for each of the periods. Since the book is organized chiefly on a geographic or provincial basis, it will be an excellent review to cut across provinces on a time basis and summarize the events over the entire continent for each period. The paleogeologic and paleotectonic maps and the brief discussion that accompanies them in Chapter 3 already do this, but no part of the text is devoted in detail to it.

The fourth type of exercise is the tracing of the geologic history of a county or a state. The commonest types of reports are those that describe the geology of an area with political boundaries, and it will serve the student as a good lesson to write a history of such a region. He will have to draw his information from several structural provinces and will find his organization, if complete, both long and complex.

2.

STRUCTURAL TERMINOLOGY

NEED OF STANDARD TERMS FOR REGIONAL STRUCTURES

The posthumous work of Schuchert (1943) is an example of the irregular use of names for the large structural features of the United States. He speaks of the Cincinnati anticline and the Cincinnati geanticline, evidently interchangeably, and the Nashville dome in the same sense as the Cincinnati anticline. McFarlan (1943), in his book on the geology of Kentucky, defines the Cincinnati arch as a major structure which includes the Jessamine dome and the Nashville dome, but in several places he refers to the arch as a dome. In Colorado the Ancestral Rockies are commonly called highlands and geanticlines, in New Mexico they are landmasses, in Texas they are uplifts and arches. The buried Nemaha "Mountains" in Oklahoma and Kansas have been called a ridge. There are a number of other terms for which no standard structural meaning has evolved. The professional geologist may not experience any difficulty or inconvenience in this loose and local application of names for the large structural features of the earth's crust, but for the student it is confusing. I have felt impelled to define and classify for his sake, because the book is addressed to him. In so doing, however, I feel at many turns there will be objections, largely on the grounds of provincial usage.

In view of the undesirability of multiplying technical words, it seems necessary to assign specific meanings to common words in their several fields of usage. For instance, the word *system* when used in stratigraphy denotes the rocks formed during a period of geologic time; when used geographically it generally signifies a group of ranges with unifying characteristics; and when used structurally it indicates a group of related joints, faults, dikes, or the like. It is probably better to give a word such as system several meanings rather than use a new word, or a less common and, perchance, a less appropriate one. The commonest usage of a term should weigh heavily in formulating a definition for it.

MEANING AND CHOICE OF TERMS FOR THIS BOOK

Arch and Dome

From 1891 to 1903 Foerste spoke of the Cincinnati uplift as an anticline, then in 1904 as a geanticline, and Schuchert continued the use of these two terms apparently interchangeably. The first mention of the terms arch and dome for the structure has not been located in the literature, but since 1900 they have been used very commonly and usually synonymously. They are the terms used both provincially and nationally most frequently today. McFarlan (1943) has distinguished the two in the sense that the Cincinnati arch is an elongate structure and includes two dome-shaped uplifts on it, the Jessamine dome and the Nashville dome, separated by a sag or saddle. Tennesseans will probably not accept the subordination of their Nashville dome to a division of the Cincinnati arch, but the principle of the distinction of arch and dome is appealing. Since

the Cincinnati and Nashville structures are the earliest of the broad, gentle uplifts studied in the United States, they probably should be taken as types, and definitions should be fashioned after their characteristics. At the completion of the present study of the uplifts and depressions of the central stable region of the United States, nothing undesirable is recognized in taking the Cincinnati and Nashville structures as types for the United States, if a little latitude in characteristics is tolerated. The terms in this report will be used as follows:

An arch is a gentle, broad uplift with an evident width of 25 to 200 miles and a length conspicuously greater than the width. The structural relief may amount to 10,000 feet or more between a bed at the top of the arch and one of similar age at the bottom of the adjacent basin, but the dip of the beds will generally not exceed 100 feet per mile. The structural relief may have been acquired in part by subsidence of the adjacent basins at a greater rate than the arch area, so that the arch may actually only at times have been an emergent landmass.

A dome is a gentle, round or elliptical uplift of arch proportions. It usually occurs along an arch and expands the arch locally. This regional structural meaning of dome must be distinguished from the usage in connection with igneous rock masses (Rice, 1940) and from the much smaller oil- and gas-producing structures such as salt domes or plugs.

Swell

Schuchert (1923) used the term swell to mean all large, domed areas within the nuclear part of the continent. Bucher (1933) defined a swell as "an essentially equidimensional uplift without connotation of size or origin." In discussing the structures of the United States the terms arch and dome are sufficient for all broad gentle uplifts, to which the term swell would generally apply, and therefore it has not been necessary to use swell in the following pages, and no attempt to define it further will be made here.

Uplift and Upwarp

Uplift and upwarp are used for a wide variety of structural elevations, and, therefore, should be reserved as noncommittal terms in regard to size, shape, internal structure, and origin. If it is desired to distinguish the two, uplift might be conceived as implying both small and large round and elongate elevations, with sharp and gentle variations; whereas upwarp would imply simply broad and gentle archings. No precedent can be cited for this distinction, but a perusal of the literature leaves me with the impression that this is the most general usage. Provincially, however, uplift may mean a rather definite type of structure. I will use the terms only in case I am in doubt about the nature of a structural elevation, or desire to use them as synonyms of structures being discussed in order to eliminate repetition.

Basin

Bucher (1933) uses the term basin in a structural sense to mean any essentially equidimensional depression without connoting size or origin, and then gives the Michigan basin as an example. Swell is his antithetical structure of basin. Since the drill in several places has extensively explored the subsurface distribution of the stratified rocks of the continent, a number of downwarps have become firmly entrenched as basins in the literature. Some embrace more than a large state, and some are of county size. Some are fairly elongate, and most all have axial directions. Some are troughlike or furrowlike. It has not proved disturbing in compiling the present review to have basin used in this loose sense, and I believe the variations in meaning will be evident to the student, so there is little urge to attach limitations to the term. The word basin is applied a thousand times each day by petroleum geologists in many variations of meaning, and it would appear unwise to attempt standardization.

Coal basins have not proved to be the same as oil basins or water basins in several places, and also the extent of the commercial materials has not coincided with the greatest thickness of the strata and, therefore, the greatest depression. It seems to me that the major geological features should govern the choice of a geographic name, rather than the distribution of an economic deposit of little relative volume.

The site of maximum subsidence during an epoch, period, or era may not coincide with that of a later one, and some confusion has resulted in the meaning of the term basin in certain areas. This is particularly true

on tectonic maps which attempt to show all structures evolved through three eras. I have found it desirable to think of certain basins in a restricted time as well as restricted geographic aspect, and to prepare accordingly the tectonic maps that accompany this book.

Geosyncline

According to Kay (1951):

> The term geosyncline should be restricted to a surface of regional extent subsiding through a long time while contained sedimentary and volcanic rocks are accumulating; great thickness of these rocks is almost invariably the evidence of subsidence, but not a necessary requisite. Geosynclines are prevalently linear, but non-linear depressions can have properties that are essentially geosynclinal.

Classifications of geosynclines are discussed by Kay, who takes the position that all basins having a thick sequence of sediments are one kind or another of geosyncline. However, only two geosynclinal terms will be used in this text, namely, miogeosyncline and eugeosyncline, which are the large linear basins along the margins of North America.

Miogeosyncline

A miogeosyncline is part of the great linear border geosyncline. It lies between the shelf regions of the stable interior of the continent and the outer part of the geosyncline. Its sediments are dominantly sandstone, shale, chert, limestone, and dolomite, almost free of volcanic rock.

Eugeosyncline

An eugeosyncline is the outer part of the border geosyncline and is characterized by an abundance of volcanic rock. In addition there is much graywacke, arkose, dark shale, and chert. The strata are generally altered by low-grade metamorphism.

Landmass

Landmass has no specific structural meaning unless used locally as in the Ancestral Rockies of New Mexico, for instance, where an ancient range is referred to as the Pedernal Landmass. The term usually connotes a land area whose elevation, climate, and life are the special object of study through the intermediary of the sediments derived from it, or whose changing shore lines form the basis of some paleogeographic study. The term does not usually imply size, relief, or origin, and no specific attributes will be affixed to it in this book.

Highland

In Colorado, two principal uplifts dominated the structural evolution of the area in late Paleozoic time, and they have been referred to by most writers as highlands. They are about 50 miles wide and 200 miles long and structurally were rather abrupt, asymmetrical anticlines which may have been faulted in part along their steep flanks. Except in application to the Colorado uplifts, the term is used very broadly in the United States, and no one to my knowledge has attempted to define it; nor is it necessary here to do so. It does not seem consistent, however, to say a certain *high*land was a *low*-lying area, but the statement may appropriately be made of a landmass.

Ridge

The buried Nemaha uplift of Oklahoma, Kansas, and Nebraska is generally spoken of as the Nemaha Mountains, but the term Nemaha ridge has also been used, with the implication that *ridge* has a certain structural significance. The use is almost unique to this area, as far as I know. A ridge, topographically, is generally less than 5 miles long, and its use structurally for the Nemaha Mountains, 200 miles long, is somewhat misleading. It is not necessary to use the term in the present review.

The term is used in oceanography to depict very large linear relief features on the ocean floor, such as the Mid-Atlantic Ridge (also Rise) or the Beata Ridge in the Caribbean Sea.

Geanticline

The term geanticline was proposed by Dana in 1873 (Schuchert, 1923) for "the upward bendings in the oscillations of the earth's crust—the geanticlinal waves or anticlinoria." According to Schuchert, Dana's typical example was the Cincinnati arch, though later on, Dana also included

far greater, even continental arching. Schuchert generally recognized geanticlines and geosynclines as "complementary structures," but called the land that divided the Cordilleran geosyncline during Mesozoic time into an eastern geosyncline and a western, the greatest of North American geanticlines.

Although Schuchert attempted to clarify Dana's most confused definition, he introduced contradictory thoughts, and therefore did not clarify the meaning of the term. Others have confused the meaning still more. According to Willis (1934), "a geanticline is a very large elevation of the earth's surface. The rocks of the geanticline may not be folded—may not even be stratified—and the anticlinal significance is lost." Lahee (1941) states that a "geanticline is a very extensive uplift, generally anticlinal in nature (also called a regional anticline)." He gives as examples the "Arbuckle Mountain Uplift and the Central Mineral Region of Texas," which are two greatly different kinds of tectonic elements. According to Nevin (1942) a geanticline is a "great upwarp . . . whose dimensions are measured in hundreds of miles. . . . The Ozark Mountains and the Arbuckle Uplift are true geanticlines." These, again, are dissimilar structures. Billings (1942) defines a geanticline as "the counterpart of a geosyncline, (it) is an area from which the sediments are derived. The geanticline that lay southeast of the Appalachian geosyncline is known as Appalachia." In the *Dictionary of Geologic Terms* (Rice, 1940) a geanticline is "a large, broad, and usually very gentle anticline, commonly many miles in width."

Most of these definitions are widely divergent, and the examples are structures of contrasting size, composition, history, and relation to the central stable interior of the continent. Some of the definitions are synonymous with terms already defined, such as arch, dome, and landmass.

The confusion in American literature is paralleled by the European. Brouwer (1925) of Holland says that a geanticline is a major uplift of island arc size, complementary to the geosyncline. Collet (1927), following Argand (1916), defines a geanticline as an anticlinal ridge that appears on the bottom of a geosyncline and expresses itself as a land barrier between the seas of the geosyncline. It is at first a long, narrow anticline of considerable size and later evolves into a great nappe. Whether the development of a nappe is necessary to demonstrate a true geanticline is not stated or implied. Most Alpine geologists, it is my impression, follow the usage of Collet.

King (1937) exemplifies the Alpine usage in his treatise of the evolution of the Marathon system. A structure in west-central Nevada that rose out of the Paleozoic Cordilleran geosyncline is called a geanticline by Nolan (1928). I have decided to follow the specific usage of Collet, King, and Nolan and will denote a geanticline as a large, elongate, anticlinal fold that develops in the sediments of a geosyncline. It is not a geanticline if an uplift in the foreland or shelf area. Two or more geanticlines may develop at the same time or following each other in a great geosyncline. After the early anticlinal uplift, the great fold usually becomes a complex anticlinorium, several imbricate thrust sheets, or a nappe. It may be largely submarine, and suffer little erosion.

Range

The synonymous use of the terms highland, landmass, mountains, uplift, arch, and geanticline, all to describe uplifts of the Ancestral Rockies and Wichita systems with fairly similar size and shape, poses a difficult problem, especially because of the provincial nature of the usage. It is so commonplace to say Electra arch and Uncompahgre highland that a change of name of one or both is not easily accepted by all. For the sake of the student who is trying to get an understanding of the rather complex, regional, structural relations of the continent, uniformity of meaning is desirable. Many geologists long since out of school recognize the need. It is a matter of clear composition.

Nearly all the structural features to which the names highland, landmass, uplift, arch, and geanticline in the Ancestral Rockies and Wichita systems have been attached are the size of a range like the Bighorn, the Uinta, or the Selkirk ranges. It seems, therefore, that the word range would be very expressive of the sharp and linear, now buried or nearly buried, late Paleozoic uplifts of the interior part of the United States.

Geographers have agreed on the usage of range and mountain system as follows: A range is a mountain mass within limits the size of the Bighorns or the Selkirks, and a number of these ranges with certain unifying

features in the region constitute a mountain system. In working over the structural features of North America I find that the divisions of the major structural provinces can fittingly be called ranges and that many of the major provinces themselves, systems. Range, therefore, will be used to denote a sharp uplift about 10 to 75 miles wide and 50 to 200 miles long. Commonly the structure is an asymmetrical anticline. In some, the steep flank has broken into a high-angle fault or a thrust. Others may consist of several folds or even thrust slices. Probably all ranges that were eventually buried suffered considerable erosion beforehand.

Platform and Shelf

The terms platform and shelf in a structural sense are logically used by King (1942a) in the Permian area of west Texas and southeastern New Mexico. There previously existing range-sized uplifts were buried, and as sediments continued to accumulate, the adjacent basins were depressed more than the old uplifts, so that although sediments accumulated on the uplifts themselves, broad anticlinal structures developed over them. These are called platforms. Beyond the basins, shallow seas existed, but the crust subsided much more slowly there than in the basins, and a much thinner deposit of sediment accumulated. These are called shelves. A platform is similar to a shelf in regard to thickness of sediments on it, but much more restricted in size and bounded on the two sides by basins. This is the sense in which the terms will be used in the following pages.

Welt and Furrow

Bucher (1933) defines welt and furrow as crustal elevations and depressions that show a distinct linear development. No special size or origin is implied. A welt may be as large as a great deformed geosyncline; viz., note Bucher's reference to Hobbs's phrase, "the gigantic welt of the Himalayas." In Bucher's analysis of the deformation of the crust on a world-wide scale, he needed these noncommittal terms, but in the present attempt to picture the structural evolution of the North American continent, the names do not seem necessary, and they will not be used.

Hinterland and Foreland

Hinterland and foreland are terms introduced by the European geologists to distinguish the landmass or resistant elements of the earth's crust on either side of an orogenic belt. In the Alps great, intricately folded masses of sediments of the geosyncline, plus injected rock, moved northward many miles. The north stable land toward which they were moved is called the foreland, and the landmass south of the geosyncline is called the hinterland. In the main, the great thrust sheets of the Appalachian and Rocky Mountain orogenic belt have overridden toward the interior stable part of the continent, and this (at least the parts adjacent to the orogenic belts) has generally been called the foreland. The landmasses or borderlands on the oceanward side have been referred to as the hinterlands. It is apparent that confusion must arise in the use of the terms when some thrust sheets have overridden toward the oceans and when, perhaps, no great, stable borderland existed. Some geologists also contended that outward from the continent is the foreland. As for usage in this book, foreland will mean the part of the stable interior adjacent to a marginal orogenic belt, and lands oceanward of a marginal trough of sedimentation, created by previous orogeny and from which sediments were derived will be called the hinterland.

TERMS FOR STRUCTURAL DISTURBANCES

Revolution and Synonyms

The term revolution is deeply intrenched in geologic literature, although a number of authors, both here and abroad, have avoided its use, and one has recommended its abandonment (Spieker, 1946).

Schuchert's (1924) definition of a revolution is more complete than any found, and characterizes many usages of the term.

> Near the close of the eras . . . occur the most extensive times of mountain making, . . . These times of major diastrophism are the *critical periods* or *revolutions* in the history of the earth, and they divide, as it were, the book of geologic time into chapters. The critical periods are marked by the following features:

1. By wide-spread deformation of the earth's crust, transmitted from place to place. This leads to the elevation of many and widely separated mountain ranges, . . . Each revolution . . . is named after one of the prominent mountain ranges formed at the time designated, for example, Laramide and Appalachian revolutions.
2. By wide-spread changes in the physical geography . . .
3. By marked and wide-spread destruction of the previously dominant, prosperous, and highly specialized organic types.
4. By marked evolution of new, dominant, organic types out of the small-sized and less specialized stocks, and by the development of hordes of new species.

With revolutions reserved to close eras, Schuchert used the term *disturbance* to terminate periods. Thus the crustal movements at the close of the Devonian period in New England and Acadia would be called the Acadian (Schickshockian) disturbance.

In light of recent research, certain disturbances are known to have occurred within periods, and three (Taconic, Acadian, Nevadan) are equal or exceed in size and certainly exceed in intensity the Appalachian (as orthodoxly known) and the Laramide revolutions. In the Alps, the diastrophic history is followed from the middle Carboniferous to the close of the Oligocene, and it seems difficult to apply the term revolution in Schuchert's sense. The great paroxysms in which the nappes were formed occurred in middle Oligocene time, and to these and all other deformations of early Tertiary time, Argand (1916) applies the name Alpine cycle. Thus he speaks as follows: "The regime of deformation of Asia during the Alpine cycle, . . . etc." (1922). He refers to the Hercynian cycle and the Caledonian cycle, apparently in the same general way as others do with the words orogeny, revolution, disturbance, and phase.

Bucher (1933) adapts the term revolution to his own nomenclature and theory by the following: ". . . the juxtaposition of the high welt and the deep sediment-filled furrow leads to the violent deformation traditionally known as 'revolutions.' "

Before deciding what terms or classification to use in this book, a few other words need to be discussed. The terms orogeny and epeirogeny, according to Gilbert (1890) are *processes* of deformation. He defined orogeny as the process of mountain building, and epeirogeny as the process of continental displacement to form the large swells and basins. The two processes collectively he called diastrophism. Orogenic structures, according to Stille (1924) are visible to the eye, such as faults, folds, and thrusts; whereas epeirogenic structures are so gentle that dips are scarcely noticeable, and are due to broad warping. The usage in America today is fairly uniform in the respect that orogenic movement is of the nature of folding, thrusting, and block faulting or rifting, and for the most part takes place in the geosynclinal belts. Epeirogenic movement is vertical, of gentle nature, and affects regional parts of the crust. The arches, domes, and large basins of the central stable region of the continent are examples of epeirogenic movements, and the interruption of cycles of erosion in the deformed geosynclinal belts by elevation is an example of epeirogenic movements in the marginal and older orogenic belts. It is in this sense that the terms will be used in this book.

A point that is confusing is the interchangeable use in our literature of *orogeny* and *revolution*. It would seem from Gilbert's early usage that orogeny is a *process*, and to say *Appalachian orogeny* would be to focus attention on the processes of deformation in the geosyncline—to emphasize the mechanical relations. On the other hand, to say *Appalachian revolution* would be to broaden one's vista structurally to the events in the hinterland and the foreland as well as in the geosyncline, and to include the climates and changes in the organic world. Current usage of the term orogeny is also often synonymous simply with crustal disturbance. Angular unconformities and coarse, thick, basal conglomerates are commonly the evidence of orogenies, and the orogenies are given names such as the Diablan, Santa Lucian, and early Laramide. Before deciding on definite usages of the terms, it is best to consider their time and geographic limits.

Phase

The term phase has been used structurally as well as stratigraphically. In nearly all structural uses it is a division, either spatial or time, of a revolution. For instance, Argand (1922) in explaining his tectonic map of Asia says, ". . . we have concluded . . . that a classification of the elements (shows) only the age of the principal folding . . . neglecting the phases but retaining the orogenic cycles." And again, ". . . all the phases

of all the orogenic cycles that have affected each part of the country, etc." Collet (1927) uses the word phase as a tectonic unit of the Middle Oligocene orogenic paroxysms of the Alps, viz., the St. Bernard phase, the Dent Blanc phase, the Monte Rosa phase, the phase of Adriatic subsidence, and the phase Insurbrienne. This usage emphasizes the mass and spatial aspect because all the nappes mentioned evolved within a short time—a succession of events is not implied. On the other hand, van Waterschoot van der Gracht (1931) uses the term more in a time aspect in describing the structural relations in the Mid-Continent area, for he designates the successive episodes of disturbance as the early Wichita phase, the late Wichita phase (early Pennsylvanian), and the Arbuckle phase (late Pennsylvanian).

Others terms such as epoch, stage, and impulse, have been used but to a lesser extent than phase.

CLASSIFICATION USED FOR CRUSTAL DISTURBANCES

Revolution

If revolutions are chapters of diastrophism in earth history, it is clear that they have both time and spatial aspects. To say they terminate the great eras of time reflects the state of advancement of the science 45 year ago. Most of the time divisions were originally set apart by unconformities, and early became more or less fixed by the fossil content of formations between the unconformities at the type localities. Since then, evidence of many new and important disturbances has been discovered within the periods and eras thus set apart. Crustal deformation has come to be known *not* as a repetition of pulsations that occurred precisely at the close of periods and eras, but as developmental sequences of deformational events which frequently occurred over protracted periods of time with shifting scenes of activity.

A revolution will be considered to encompass the deformational events of the hinterland, the geosyncline, and the foreland, and to include both orogenic and epeirogenic processes. Setting time limits is an arbitrary procedure, and in doing so one must be mindful of usage which will help determine the best limits of the revolution in question.

System

The major structural divisions of revolutions will be called systems. A system is thus primarily a spatial division and is determined by a unity of the structural features in it, such as the folds and thrusts of a geosyncline in contrast to the basins, shelves, and arches of the foreland, or by isolation of a somewhat similar structural assemblage from another by younger overlapping deposits, such as separate the Ouachita Mountains from the Marathon Mountains.

As far as noted, systems have been named after the outstanding range or geographic feature in the division. This precedent will be followed structurally where possible, but some exceptions seem necessary. For instance, in organizing the structures of the central stable region of the United States the area proved so large that no one geographic name seemed suitable for the greatest arch, so its outstanding structural character was used, namely, the Transcontinental Arch.

Phase

Each system has its developmental history, and the structural events of this history will be called phases. Although the types and extent of the structures developed will be considered part of the phase, emphasis is laid on the time aspect. It may be necessary to consider as phases two contemporaneously evolving parts of a system, but in organizing structural elements of the continent I have not run into this difficulty.

In the Alps, the phases have been given geographic names, and the practice was followed in this country by Van der Gracht (1931), who discussed the Wichita and Arbuckle phases of the Wichita system. Since in this book the emphasis will be placed on time, I have concluded that time names will be most meaningful. For instance, if it is written, *the early Pennsylvanian phase of the Ancestral Rockies system*, the student cannot miss the intended meaning; but if *the Wichita phase of the Ancestral Rockies* appears, the student may be confused if he has not read the chapter on the Wichita system.

Time names can be inappropriate only where the time of the disturbance is not accurately known and future research shows the designation

wrong. A geographic name avoids this difficulty, it is true, but for the most part stratigraphy has advanced to the point, in the United States at least, that the times of the deformation are fairly accurately known and not likely to be changed much in the future. The advantage to the student weighs so heavily against the possible chance of error that time names for the phases will be used.

Most of the chapters deal with systems and their phases. Such organization seems adequate to explain the structural evolution of the continent. Originally it was planned to organize the book according to revolutions, but setting limits led to many difficulties, and the idea was abandoned. As a result, the concept of revolution is not emphasized.

Orogeny

With the decision reached to divide the great deformational belts into mountain systems, and to treat the several episodes of deformation of each system as phases, the proper usage of the term orogeny seemed clear: each phase is an orogeny. Thus we speak of the "Late Cretaceous and Early Tertiary Rocky Mountain systems," and for one of these systems, the Central Rockies, we note its episodes of deformation, namely, the Montana phase, the Paleocene phase, and the Eocene phase. These phases are commonly the orogenies, which respectively would be the early Laramide orogeny, middle Laramide orogeny, and late Laramide orogeny. See table of contents for the various orogenies recognized in North America.

An orogeny should be given a geographic name, like a formation, and if the time of deformation is found to be earlier or later than previously recognized on the basis of later research, then the name remains the same, but a somewhat different age is assigned it.

An orogeny should not be limited to a phase of folding and thrusting, but should include all forms of diastrophism, according to Billings (1960).

3.

RÉSUMÉ OF STRUCTURAL GEOLOGY OF NORTH AMERICA

MAJOR TECTONIC DIVISIONS

Canadian Shield

The Canadian Shield has been the great stable portion of the North American continent since Proterozoic time. It consists of Precambrian rock except along the southern margin of Hudson Bay, where Ordovician, Silurian, and Devonian strata, about 1000 feet thick, occur and probably continue northward under much of the bay. Small outliers of Paleozoic strata, fossil affinities, and the absence of shore facies in many places indicate that the Paleozoic formations were once much more widespread over the shield than now, and that they have been stripped off by a long interval of erosion during the Mesozoic and Cenozoic eras.

Hudson Bay is an epeiric sea of fairly modern time.

Central Stable Region

The Central Stable Region consists of a foundation of Precambrian crystalline rock, which is a continuation of the Canadian Shield southward and westward, and a veneer of sedimentary rock. The veneer varies greatly in thickness from place to place, and several broad basins, arches, and domes are present. A number of unconformities attest the rise of the arches and their erosion, and of great transgressions and overlaps. For the most part the strata have only gentle dips, and aside from the slow and prolonged vertical movements that created the basins, arches, and domes, the geologic province properly deserves the name, the Central Stable Region. It and the Canadian Shield compose the great stable interior of the continent.

The arches and basins developed chiefly in the Paleozoic era, but later, during the Mesozoic and Tertiary, vast amounts of clastic sediments from the evolving Cordilleran mountain systems were spread eastward over the Paleozoic strata beyond the Missouri River as far as Lake Superior.

In the southwestern corner of the Central Stable Region a system of ranges was elevated in Pennsylvanian time, and then during the late Pennsylvanian, Permian, and Mesozoic it was largely buried. The ranges are known as the Ancestral Rockies in Colorado and New Mexico, and as the Wichita Mountain system in Kansas, Oklahoma, and Texas. The Late Cretaceous and Early Tertiary Laramide structures were partly superposed on the Ancestral Rockies in Colorado and New Mexico.

Orogenic Belts of the Atlantic Margin

The Paleozoic orogenic belts bound effectively the southern, as well as the eastern, margin of the continent. The major belt is known as the Appalachian, and it consists of an inner folded and thrust-faulted division from Alabama to New York, and a metamorphosed and intruded division from Alabama to Newfoundland. One major orogeny occurred in the

inner belt, and this in late Paleozoic time. Several orogenies beset the outer belt: the earliest one of significance occurred at the close of the Ordovician, the major one during the Late Devonian and the last one in Pennsylvanian and Permian time. The Carboniferous orogenic belt in the outer crystalline division is recognized on the north along the eastern margin of New England, the Maritime Provinces, and Newfoundland.

Volcanic rocks and great batholiths are important components of the crystalline division of the Appalachian orogenic belt, but the inner folded and thrust-faulted belt is comparatively free of them. Both divisions are made up of very thick sedimentary sequences which are characterized as geosynclinal, in contrast to generally thinner sequences in the Central Stable Region.

The orogenic belt bordering the southern margin of the stable interior is mostly concealed by overlapping coastal plain deposits. Where exposed, as in the Ouachita Mountains of Arkansas and eastern Oklahoma, the Arbuckle Mountains of south central Oklahoma, and the Marathon Mountains of western Texas, it is a folded and thrust-faulted complex, somewhat similar to the inner Appalachian division. The crystalline division, if it parallels the inner noncrystalline division, is nowhere exposed, but deep wells through the coastal plain deposits have penetrated low-grade metamorphic rocks.

Orogenic Belts of the Pacific Margin

The great complex of orogenic belts along the Pacific margin of the continent has evolved through a very long time. The oldest strata recognized from their fossils are Ordovician, and deformed strata of Pleistocene age mark the belt in places from Mexico to Alaska. In Paleozoic time, the Pacific margin of the continent was a volcanic archipelago in outward appearance and internally a belt of deformation and intrusion. The Permian, Triassic, and Early and Middle Jurassic were times of excessive volcanism, and represent a continuation of essentially the same Paleozoic conditions well into the Mesozoic. In Late Jurassic and early Late Cretaceous time, intense folding and batholithic intrusions (Nevadan orogeny) occurred which are now characteristic of large parts of the Coast Range of British Columbia, the ranges along the international border in British Columbia, Washington, and Idaho, the Klamath Mountains of southwestern Oregon and northern California, the Sierra Nevada of California, and the Sierra of Baja California. The same Nevadan elements may also continue into southern Mexico and eastward through Central America.

Following the orogeny, in California at least, a new trough of accumulation and a new volcanic archipelago formed outside the Nevadan belt, and a complex history of deformation and sedimentation carries down through the Cretaceous and Tertiary to the present, to result in the Coast Ranges of Washington, Oregon, and California.

Orogenic Belts of the Rocky Mountains

During the complex and long orogenic history of the Pacific margin, the adjacent zone inward was one of gentle subsidence and sediment accumulation, comparatively free of volcanic materials, during the Paleozoic. By Triassic time, the troughs of deposition along the Pacific had become effectively separated by a medial, linear uplift from those in the Rocky Mountain area, and in the Mesozoic much coarse debris came from the uplift or geanticline and filled the basins in eastern British Columbia, western Alberta, Idaho, western Wyoming, central Utah, and southern Nevada. Orogeny from place to place along the eastern margin of the geanticline cast several floods of conglomerate eastward during the Cretaceous.

The Paleozoic and all the Mesozoic sediments except the Upper Cretaceous of the Rocky Mountains may be divided into thick geosynclinal facies on the west and fairly thin shelf facies on the east. The line dividing the two lies approximately along the west side of the Colorado Plateau and thence runs northward through western Wyoming and Montana to western Alberta. The shelf facies were part of the Central Stable Region until the Late Cretaceous and Early Tertiary (Laramide) orogeny in whose belts both geosynclinal and shelf facies were deformed. The western division of the Laramide belt (in the miogeosyncline) is characterized by folds, thrusts, and numerous small stocks. The eastern

Laramide division extended through the shelf region of central and eastern Wyoming, central Colorado, eastern Utah, and central New Mexico, and is characterized by large, elliptical uplifts.

The Laramide belt of orogeny extends southward through Mexico, where thick sediments of the Mexican geosyncline of Upper Jurassic and Cretaceous age are fairly tightly folded. The same belt of orogeny is believed to veer eastward through Central America.

Following well after the Nevadan and Laramide orogenies of western North America, an episode of high-angle faulting occurred, that created the Great Basin physiographic province and gave sharp definition to many of its ranges and to those of central and western Mexico. The high-angle faults were superposed on both the Nevadan and Laramide belts; most of them are Late Tertiary in age and some are still active. A long zone of the faults extends northward from central Utah to British Columbia and probably beyond to Yukon Territory to form a belt of trenches with local relief of 3000 to 5000 feet. The faults cut the older folds and thrusts both discordantly and concordantly, and the activating forces appear deep-seated.

Coastal Plains

Following the Appalachian orogeny in Triassic time, the outer metamorphosed division was broken by a belt of high-angle faults that has been traced discontinuously from South Carolina to the Bay of Fundy, between New Brunswick and Nova Scotia. Long and narrow downfaulted basins trapped thick series of generally red clastics. The Triassic lowland of Maryland, New Jersey, and Pennsylvania, and the central lowland of Connecticut are the best known of the basins.

The eastern extent or breadth of the Appalachian orogenic system and the nature and condition of the crust that lay east of it are not known, but the continental margin had begun to subside, at least by Early Cretaceous time, if not before. The peneplained surface on the crystalline rocks has been traced eastward under a Cretaceous and Tertiary sedimentary cover to a depth of 10,000 feet, which is near the margin of the present continental shelf. Most sedimentary units of the cover dip gently and thicken like a wedge oceanward as far as they have been traced by deep drilling and by seismic traverses. The zone of Cretaceous and Tertiary overlap on the older rocks of the eastern continental margin is known as the Atlantic Coastal Plain, but because the same sediments continue out beyond the present ephemeral shore line, the submerged part belongs to the same province. Coastal plain sediments are known to exist in Georges Bank off Rhode Island and probably make up part of, or all, the shallow continental shelf to and including the Banks of Newfoundland.

The Gulf Coastal Plain is continuous with the Atlantic Coastal Plain, and counting its shallowly submerged portions, it nearly encloses the Gulf of Mexico. The oldest known sediments of its marginal areas are Permian. The Mississippi, Rio Grande, and other rivers draining the interior of the continent have deposited a great weight of sediments at their mouths and the crust has subsided along the Texas, Louisiana, and Mississippi coast to the extent of 25,000 to 30,000 feet.

Deep drilling in Florida and the Bahamas indicates that the coastal plain province extends southeastward almost to the orogenic belt of Cuba and Hispaniola.

Canadian Arctic

The Precambrian rocks of the Canadian Shield are overlapped on the north by nearly flat-lying sedimentary strata of Paleozoic age. Basins and arches are recognized in this province as in the Central Stable Region of the United States. North of the arches and basins is a fold belt developed in geosynclinal sediments. The fold belt extends across northern Greenland, northern Ellesmere Island and farther to the southwest through other islands of the Arctic Archipelago. Folding first occurred in pre-Pennsylvanian time. After erosion a voluminous sequence of Pennsylvanian to Tertiary sediments accumulated, and then these were somewhat folded in Tertiary time. A narrow Tertiary coastal plain is terminated on the north by the Arctic Ocean basin.

Alaska

Alaska continues the broad and complex western cordillera across to Asia, and has had basically the same history but with variations and singular details.

Meaning of Colors on Tectonic Maps
(Plates 2-15)

BLUE Denotes regions of accumulations of sediments. Contours indicate thickness of sediments and thus, approximately, the amount of subsidence. Thickness figures indicate thousands of feet.

GREEN Denotes ocean basins; i.e., regions underlain by oceanic crust.

ORANGE Denotes significant deformation. Where sediments have been deformed during the period in which they were deposited such has been printed on the blue.

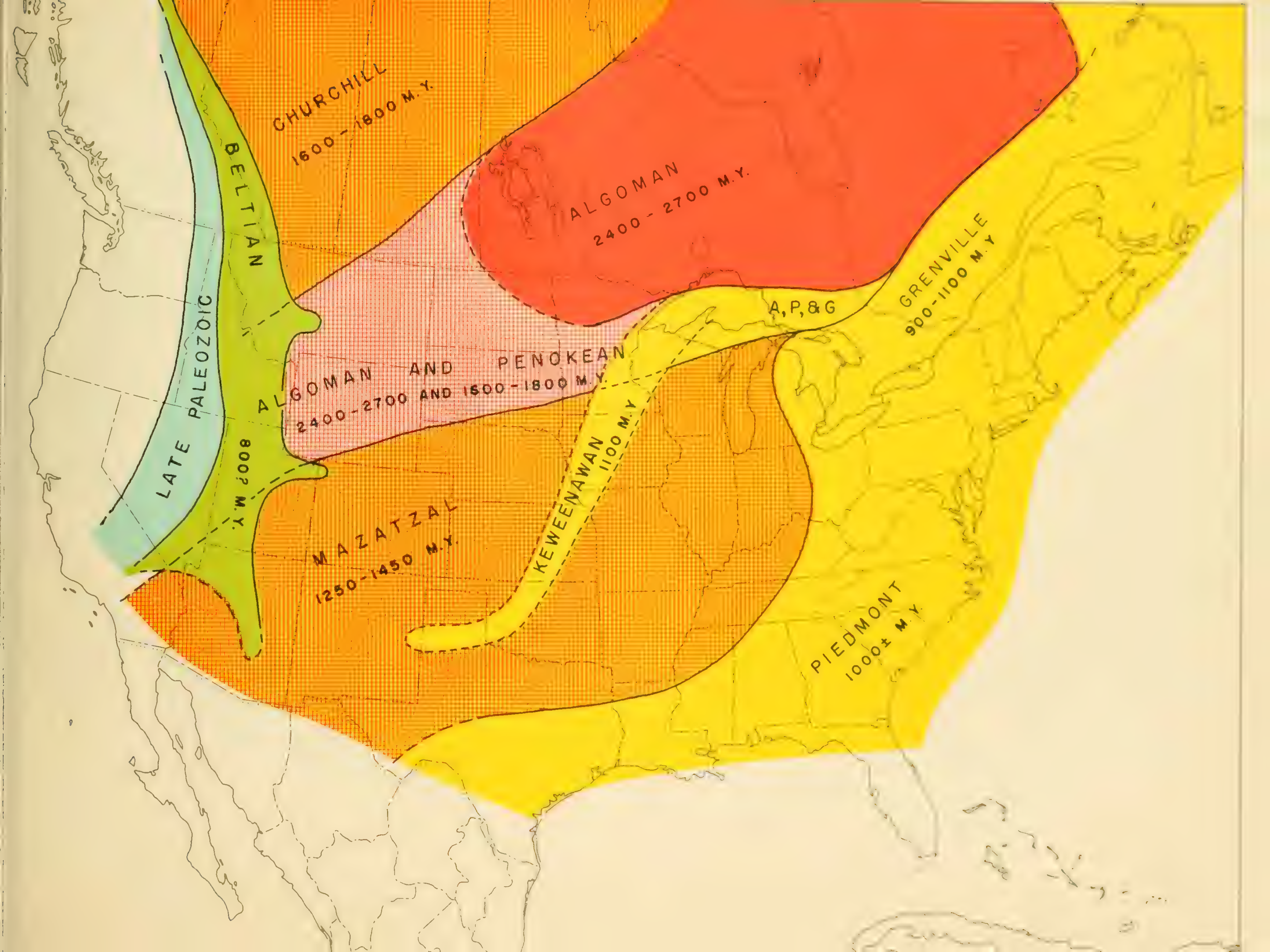
RED Denotes belts of batholithic intrusion and appreciable metamorphism on all Plates except 1, 14, and 15. On Plate 1 various intensities of red plus orange and yellow denote orogenic belts of different ages. On Plates 14 and 15, red denotes igneous rock, chiefly volcanic.

YELLOW Denotes regions of comparative stability of the earth's crust. It includes on some maps regions of broad and gentle uplift (Plate 1 excepted).

PLATE 1

Precambrian Orogenic Belts

Position of belts older than the Beltian is determined principally by absolute isotope ages. A, P, and G are dates of Algoman, Penokean, and Grenville orogenies, respectively.

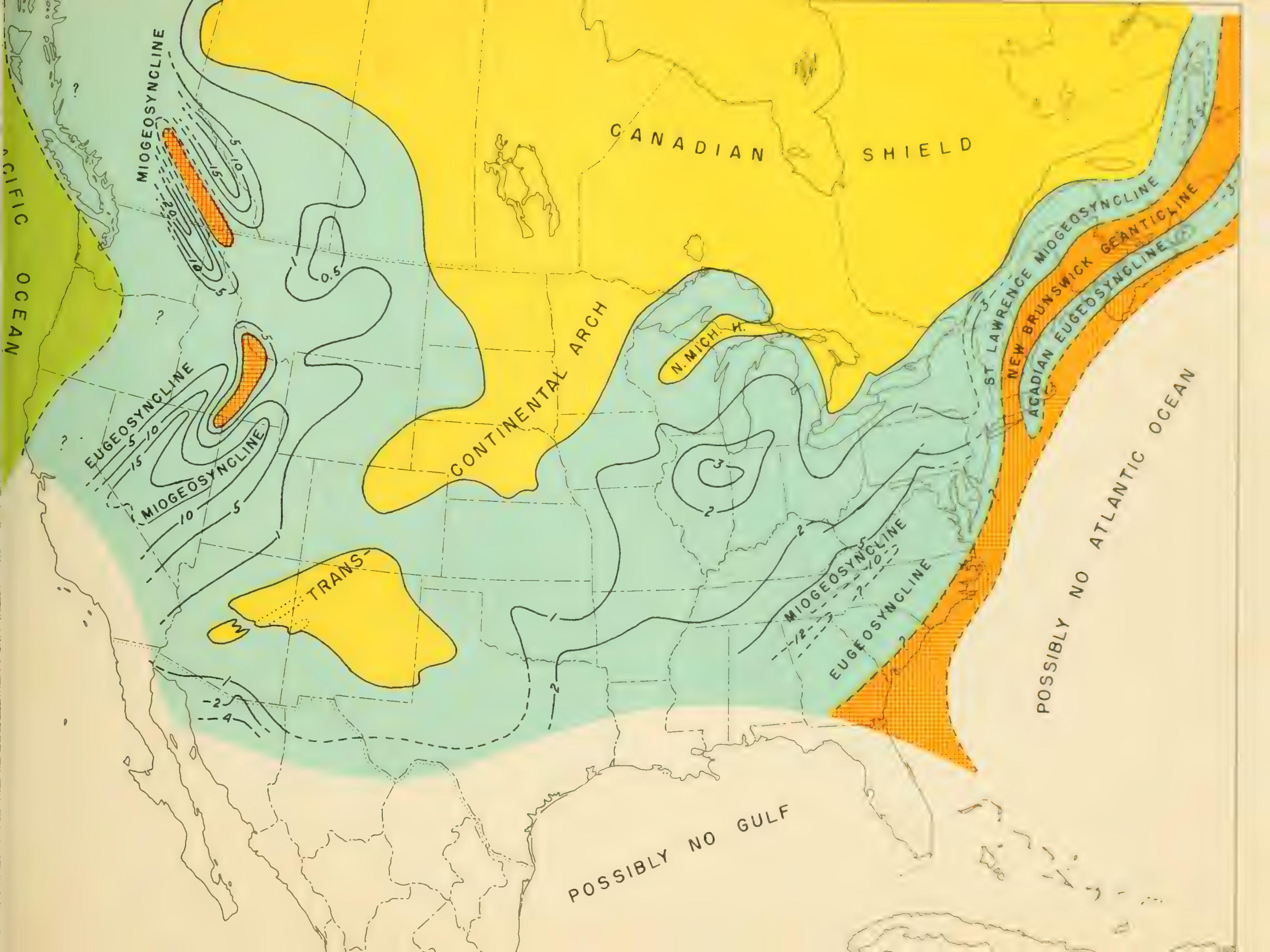


CHURCHILL
1600 - 1800 M.Y.
BELTIAN
800? M.Y.
LATE PALEOZOIC
ALGOMAN
2400 - 2700 M.Y.
ALGOMAN AND PENOKEAN
2400-2700 AND 1600-1800 M.Y.
A, P, & G
GRENVILLE
900-1100 M.Y.
MAZATZAL
1250-1450 M.Y.
KEWEENAWAN
1100 M.Y.
PIEDMONT
1000± M.Y.

PLATE 2

Cambrian Tectonic Map

Upper Cambrian seas were probably more widespread in the Transcontinental Arch region than shown; the strata have been eroded away there. The Cambrian beds of eastern Newfoundland, although evidently in the Acadian trough, are mostly miogeosynclinal in lithology.

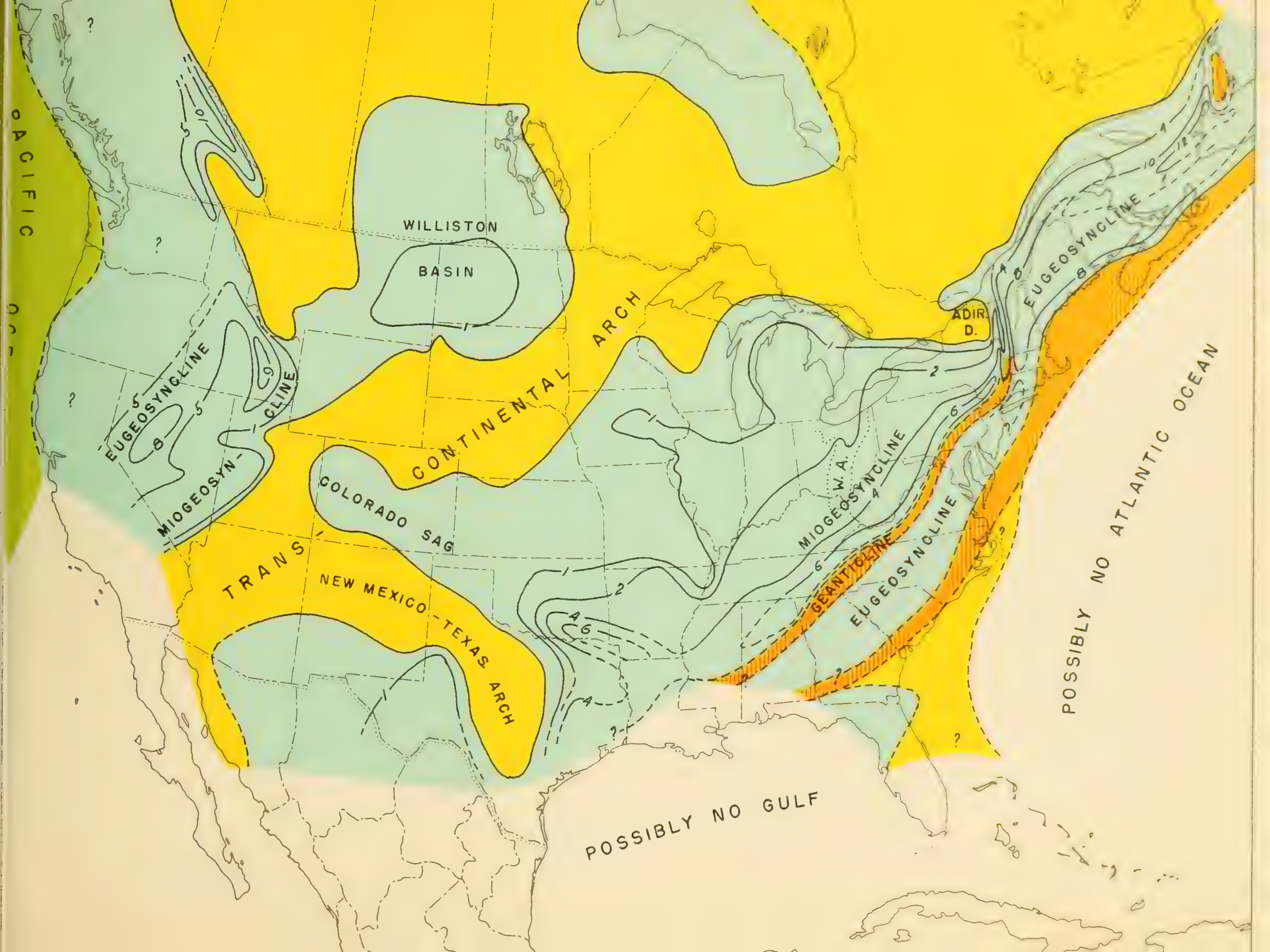


CANADIAN SHIELD
OCEAN
MIOGEOSYNCLINE
EUGEOSYNCLINE
MIOGEOSYNCLINE
CONTINENTAL ARCH
N. MICH. H.
TRANS-
ST LAWRENCE MIOGEOSYNCLINE
NEW BRUNSWICK GEANTICLINE
ACADIAN EUGEOSYNCLINE
MIOGEOSYNCLINE
EUGEOSYNCLINE
POSSIBLY NO ATLANTIC OCEAN
POSSIBLY NO GULF

PLATE 3

Ordovician Tectonic Map

Westward thrusting occurred at the close of the Ordovician in eastern New York against the Adirondack dome. Some ultrabasic intrusions may have been emplaced in the Maritime Provinces and Newfoundland at the close of the Ordovician. W. A. Waverly arch of Early Ordovician time.

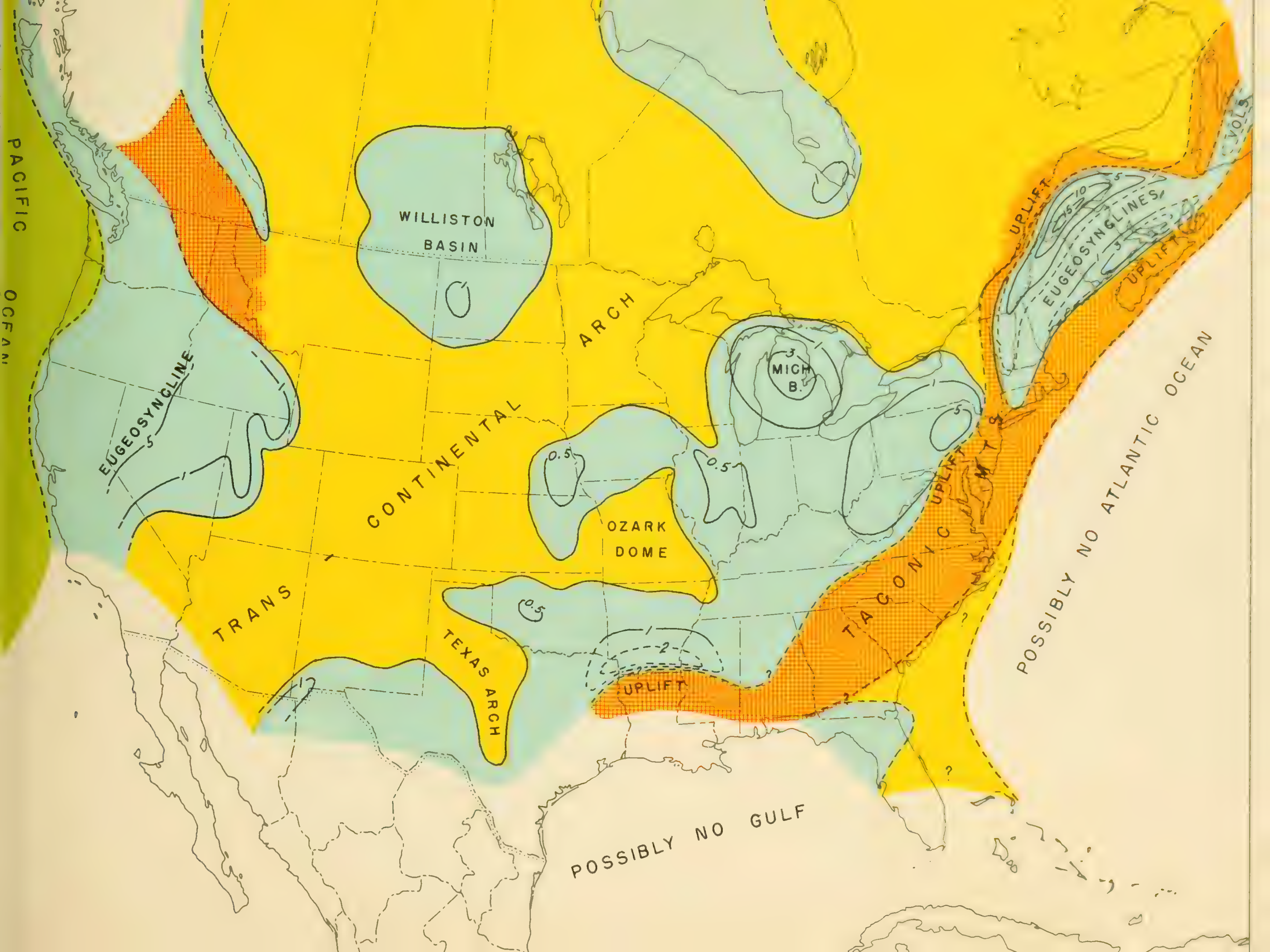


PACIFIC
WILLISTON
BASIN
EUGEOSYNCLINE
MIOGEOSYN-
CLINE
CONTINENTAL ARCH
TRANS
COLORADO SAG
NEW MEXICO-TEXAS ARCH
ADIR.
D.
EUGEOSYNCLINE
W. A.
MIOGEOSYNCLINE
GEANTICLINE
EUGEOSYNCLINE
POSSIBLY NO ATLANTIC OCEAN
POSSIBLY NO GULF

PLATE 4

Silurian Tectonic Map

The Atlantic Ocean and Gulf of Mexico are left uncolored because accumulating evidence suggests that North America was once attached to and part of a single great continent which cracked and drifted apart. The spreading apart is presumed to have brought these ocean basins into existence, starting in late Paleozoic time.



PACIFIC
WILLISTON BASIN
CONTINENTAL ARCH
TRANS
EUGEOSYNCLINE
OZARK DOME
TEXAS ARCH
MICH. B.
UPLIFT
TACONIC MTS.
UPLIFT
EUGEOSYNCLINES
UPLIFT
UPLIFT
VOLS
POSSIBLY NO ATLANTIC OCEAN
POSSIBLY NO GULF

PLATE 5

Devonian Tectonic Map

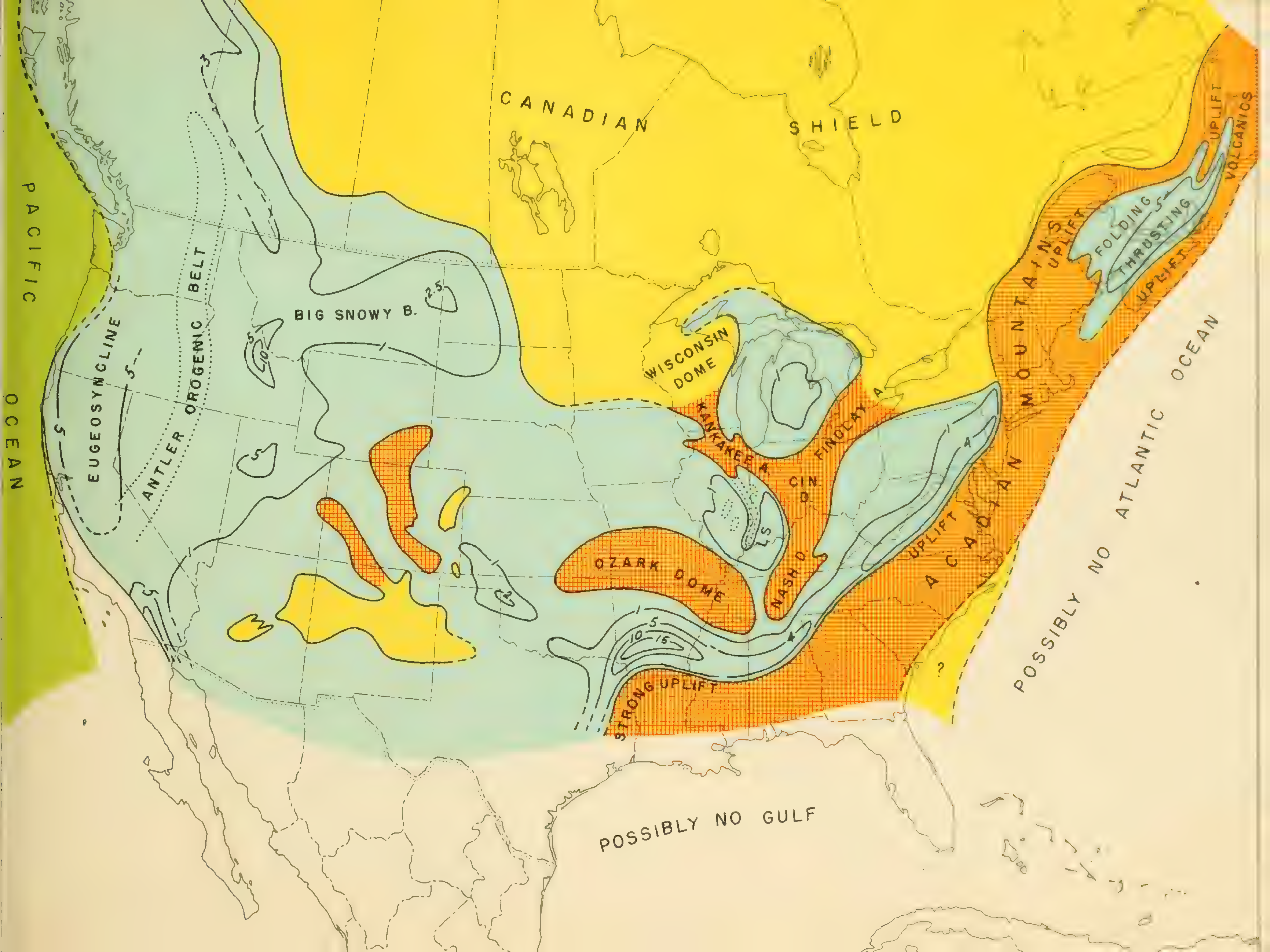
The eugeosynclinal regions in Acadian orogenic belt preceded the orogeny; their sediments were intensely deformed and invaded by the large batholiths, and hence are not shown in blue.


PACIFIC OCEAN
ALBERTA BASIN
WILLISTON BASIN
SHALLOW SEAS AT TIMES ACROSS THIS ARCH
EUGEOSYNCLINE
ANTLER OROGENIC BELT
UTAH - WYOMING SHELF
CONTINENTAL ARCH
MICHIGAN BASIN
EARLY DEV. EUGEOSYNCLINE
EUGEOSYNCLINE
ACADIAN BATHOLITHIC AND OROGENIC BELT
LATE DEV. BASIN
ILL. IND. KY. B.
CIN. D.
NASH. D.
OZARK DOME
ELLIS A.
COLORADO SAG
TRANS - NEW MEXICO - TEXAS ARCH
POSSIBLY NO ATLANTIC OCEAN
POSSIBLY NO GULF

PLATE 6

Mississippian Tectonic Map

Stanley, Jack Fork, and Johns Valley clastics of Ouachita Mountains are shown as a Mississippian basin. They may be in part Pennsylvanian. LS means La Salle anticlinal belt. It rose gently, was eroded, and buried before the Mississippian period ended. Orange here indicates areas of orogeny, significant uplift, or mountains of an immediately prior orogeny. In the Antler orogenic belt both sedimentation and orogeny occurred.



CANADIAN SHIELD
PACIFIC OCEAN
EUGEOSYNCLINE
ANTLER OROGENIC BELT
BIG SNOWY B.
WISCONSIN DOME
KANKAKEE A.
FINDLAY A.
CIN. D.
NASH. D.
OZARK DOME
STRONG UPLIFT
UPLIFT
ACADIAN MOUNTAINS
UPLIFT
FOLDING
THRUSTING
UPLIFT
UPLIFT VOLCANICS
POSSIBLY NO ATLANTIC OCEAN
POSSIBLY NO GULF

PLATE 7

Pennsylvanian Tectonic Map

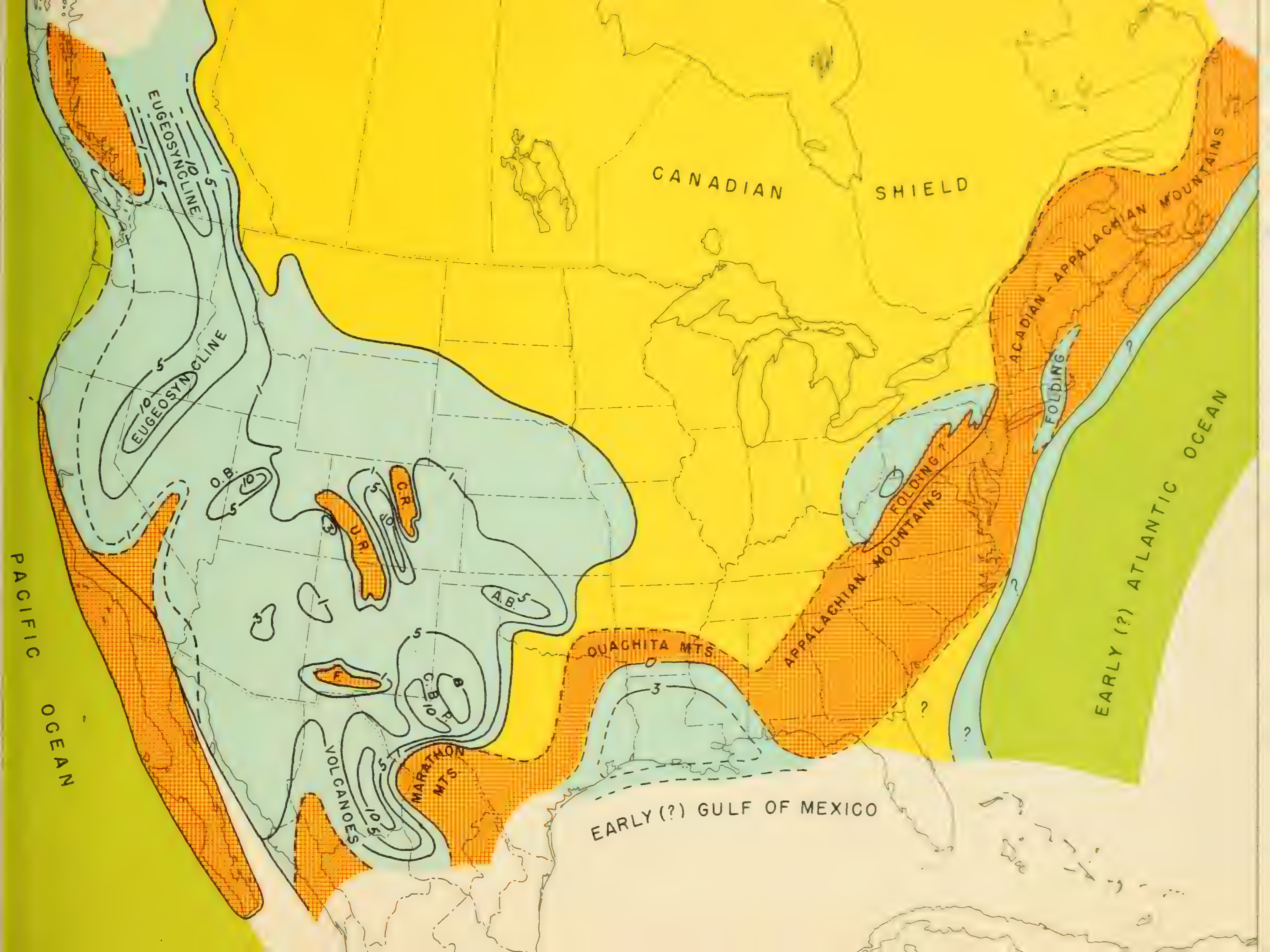
Uplifts shown by dotted lines were mostly buried by end of Pennsylvanian. The Baja California block lay several hundred miles to the southeast. A, Marathon basin; B, Fort North basin; C, Ouachita basin; D, Southern Appalachian basin; E, Central Appalachian basin; F, Diablo Range; G, Pecos Range; H, Pedernal Range; I, Zuni uplift; J, Circle Cliffs uplift; K, Emery uplift; L, Oquirrh basin; M, Central Colorado basin; N, Wood River basin; P, Ardmore basin; T, Matador Range; W, Amarillo-Wichita Range.

PACIFIC OCEAN
UPLIFT AND DEFORMATION
EUGEOSYNCLINE
FOLDING
THRUSTING
BELTIAN GEANT.
CAMBRIDGE-CEN. KAN. A.
NEMAHA R.
FOREST CITY BASIN
ILL. IND. KY. B.
COL. R.
U.R.
PARADOX B.
THRUSTING
STRONG UPLIFT
STRONG UPLIFT
ACADIAN
MOUNTAINS
FOLDING
UPLIFT
POSSIBLY NO ATLANTIC OCEAN
POSSIBLY NO GULF

PLATE 8

Permian Tectonic Map

Orange color over the Marathon, Ouachita, and Appalachian Mountains indicates the site of an orogenic belt and mountains of the previous Pennsylvanian period. Uplift probably occurred there during the Permian. U.R., Uncompahgre Range; C.R., Colorado Range; F, Florida uplift; O.B., Oquirrh basin; A.B., Anadarko basin; C.B.P., Central basin platform.



CANADIAN SHIELD
EUGEOSYNCLINE
EUGEOSYNCLINE
O.B.
C.R.
U.R.
A.B.
F.
C.B.P.
VOLCANOES
MARATHON MTS.
OUACHITA MTS.
APPALACHIAN MOUNTAINS
FOLDING ?
ACADIAN APPALACHIAN MOUNTAINS
FOLDING
PACIFIC OCEAN
EARLY (?) ATLANTIC OCEAN
EARLY (?) GULF OF MEXICO

PLATE 9

Triassic Tectonic Map

Baja California block lay several hundred miles to the southeast.

CANADIAN SHIELD
EUGEOSYNCLINE
DOWN-FAULTED BASINS
PACIFIC OCEAN
ATLANTIC OCEAN
GULF OF MEXICO
3
1
5
5
1
5
0
15
10
2
0
1
2
5
2

PLATE 10

Jurassic Tectonic Map

Baja California block lay several hundred miles to the southeast.


CANADIAN SHIELD
HILLS
HILLS
EUGEOSYNCLINE
GEANTICLINE
MEXICAN GEOSYNCLINE
PACIFIC OCEAN
ATLANTIC OCEAN
GULF OF MEXICO
25
15
5
1
0
5
1
0
1
5
4
1

PLATE 11

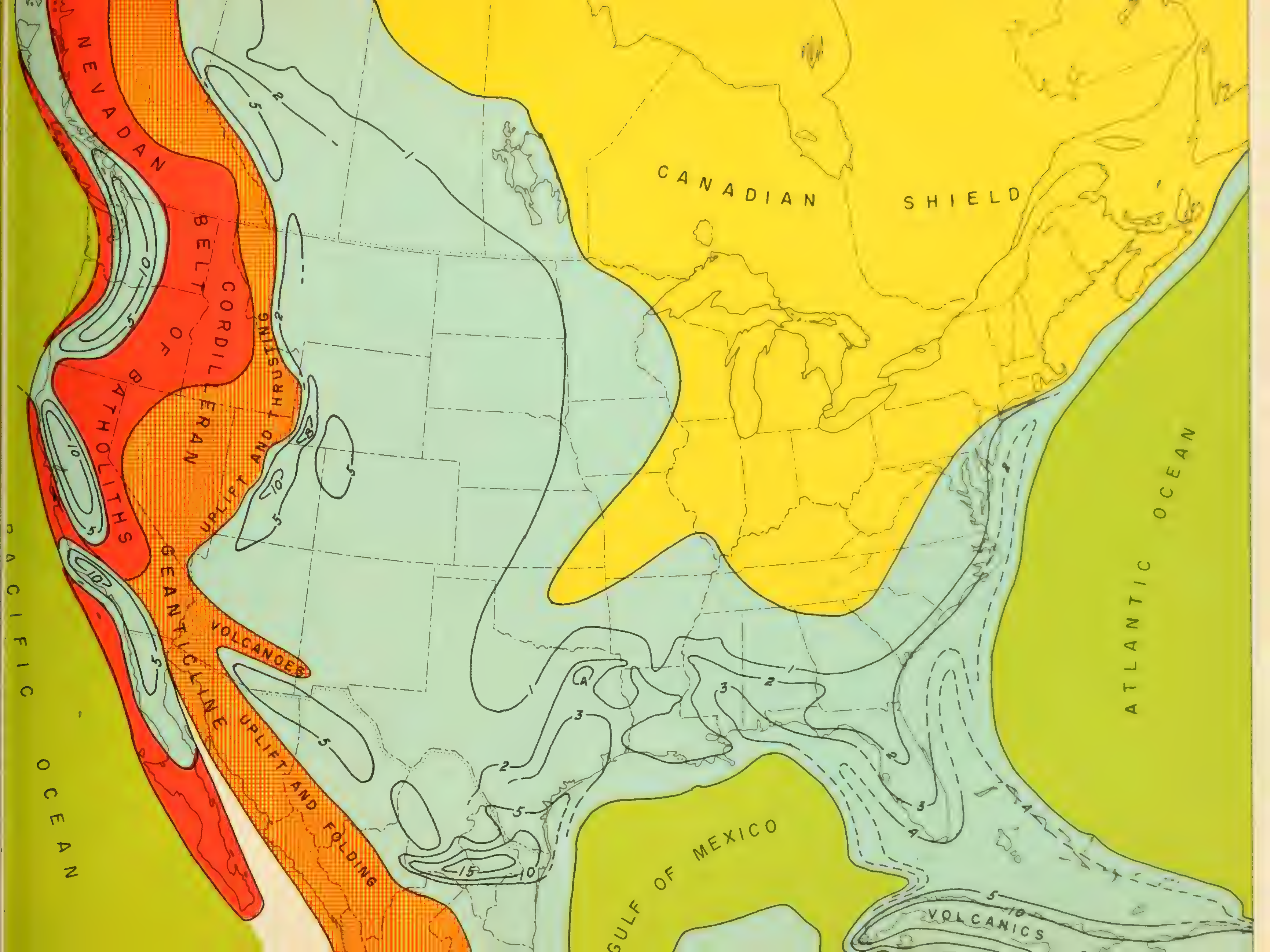
Early Cretaceous Tectonic Map


CANADIAN SHIELD
BELT OF STRONG UPLIFT
CORDILLERAN GEANTICLINE
PACIFIC OCEAN
ATLANTIC OCEAN
GULF OF MEXICO
VOLCANOES
MEXICAN GEOSYNCLINE
VOLCANICS
750
250
500
700
15
0
1
2
3
5
10
15
20
25
30

PLATE 12

Late Cretaceous Tectonic Map

Only Dakota and Colorado deposits in Rocky Mountains are represented. Montana time is shown on Plate 13. Main batholiths of Nevadan orogenic belt were intruded in very early Late Cretaceous time.



CANADIAN SHIELD
NEVADAN BELT OF BATHOLITHS
CORDILLERAN UPLIFT AND THRUSTING
GEANTICLINE UPLIFT AND FOLDING
VOLCANOES
VOLCANICS
PACIFIC OCEAN
ATLANTIC OCEAN
GULF OF MEXICO

PLATE 13

Tectonic Map of the Cretaceous-Tertiary Transition

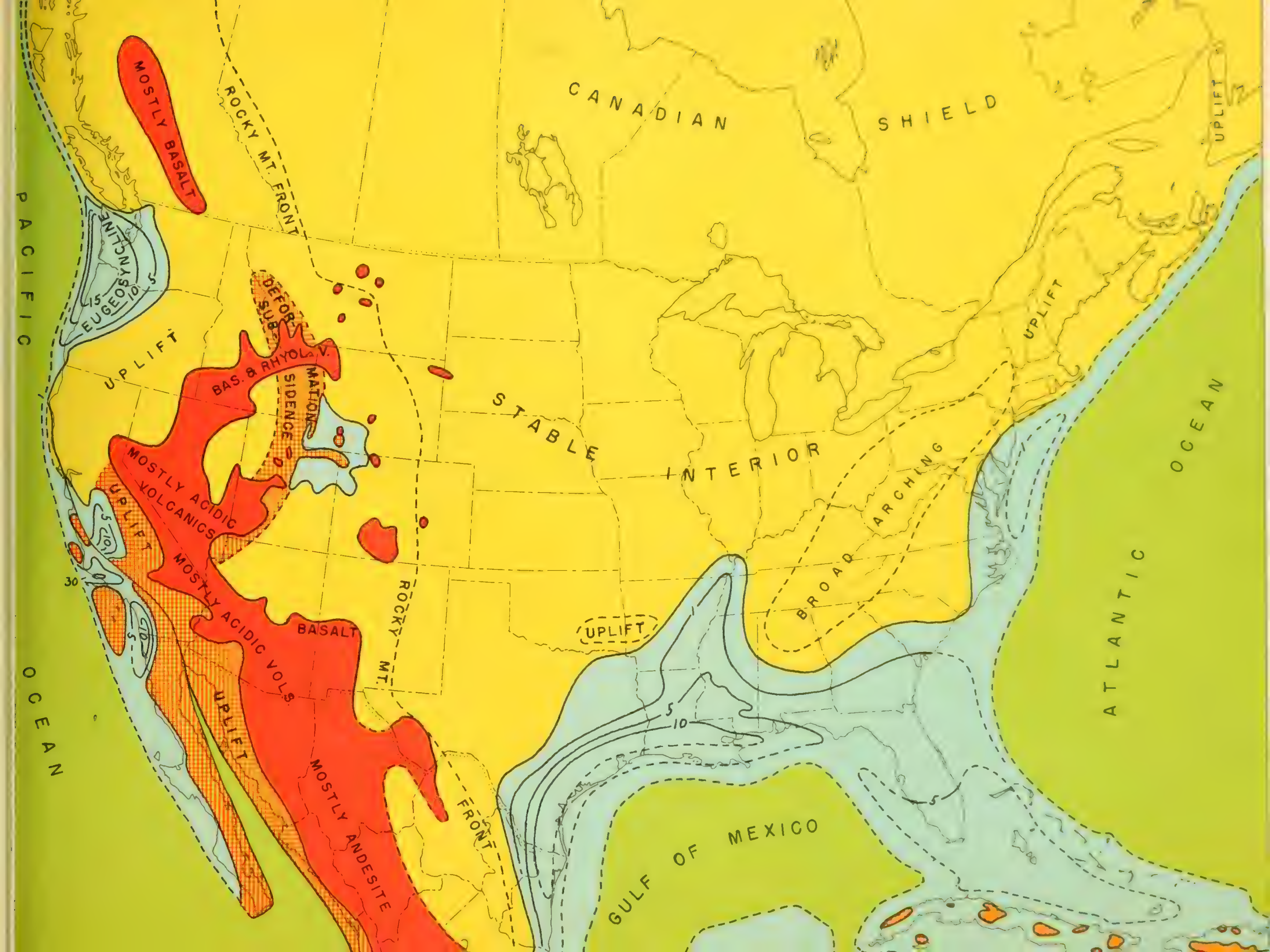
Thickness of latest Cretaceous and Early Tertiary deposits in Rocky Mountain basins not shown. For detail see Figs. 22.4, 22.5, and 22.6. The crypto-volcanic structure in Iowa is Late Cretaceous or Early Tertiary; the others are not dated but are presumed to be of the same age.

CANADIAN SHIELD
FOLD AND THRUST BELT OF THE ROCKY MOUNTAINS
CORDILLERAN GEANTICLINE
ANTICLINAL UPLIFTS OF SHELF
FOLD BELT OF MEXICO
PACIFIC OCEAN
ATLANTIC OCEAN
GULF OF MEXICO
NORTHWARD THRUSTING

PLATE 14

Early Tertiary Tectonic Map

Laramide uplifts and basins not shown except for Green River Lake in Utah, Colorado, and Wyoming, and Uinta Range, although thick Eocene deposits accumulated in most of them. The Rocky Mountain front is a result of previous orogeny. The volcanics are Eocene, Oligocene, and in places Miocene in age; most Miocene volcanics, however, are shown on Plate 15. Numerous volcanic cones, not shown, were built in the eastern Pacific and the Gulf of California.



MOSTLY BASALT
ROCKY MT. FRONT
CANADIAN
SHIELD
UPLIFT
PACIFIC
EUGEOSYNCLINE
15
10
5
DEFORMATION
SUB SIDENCE
BAS. & RHYOL. V.
UPLIFT
STABLE
INTERIOR
MOSTLY ACIDIC VOLCANICS
UPLIFT
UPLIFT
BROAD ARCHING
ATLANTIC
OCEAN
30
MOSTLY ACIDIC VOLS.
BASALT
ROCKY MT.
UPLIFT
UPLIFT
OCEAN
MOSTLY ANDESITE
FRONT
GULF OF MEXICO

PLATE 15

Late Tertiary and Quaternary Tectonic Map

Red, besides denoting volcanic rocks, shows laccolithic clusters in the Colorado Plateau. Numerous centers of volcanism throughout the Basin and Range province are not shown. The blue color extends to lines of maximum transgression of seas during the time represented by the map. Hudson Bay and St. Lawrence submergence pre-dates the post-glacial uplift. The submerged coastland of British Columbia has been uplifted 600 feet in post-glacial time. See Fig. 31.25 for regional vertical movements of the western Cordillera in late Cenozoic time.

SUBMERGENCE
900
CANADIAN
SHIELD
500
PLEISTOCENE SUBMERGENCE
UPLIFT IN POST-GLACIAL TIME
700
UPLIFT
SUBSIDENCE
ROCKY MT. TRENCH
REGIONAL
UPLIFT
PROGRESSIVELY
GREATER TO
WEST
PACIFIC
OCEAN
FISSURE
BASALT
FIELD
VENT BASALT FIELD
STRATOVOLCANOES
FOLDING
UPLIFT
TRENCHES
DOWNFAULTED
BLOCKS
D
STABLE
INTERIOR
HINGE
LINE
0
500
UPLIFT
BLOCK
FAULTING
SUBSIDENCE
UPLIFT
B
REGIONAL
UPLIFT
BROAD
ARCHING
UPLIFT
SUBSIDENCE
BLOCK
FAULTING
UPLIFT
VOLCANOES
SPLITING APART
5
10
20
5
8
12
GULF OF MEXICO

Central America

Southern Mexico, Guatemala, Honduras, El Salvador, and Nicaragua contain a belt of metamorphic rocks which sweeps from southwestern Mexico in an easterly direction to the Caribbean Sea. A belt of deformed Permian strata with Permian (?) granitic and ultrabasic intrusives makes up part of the crystalline complex. A fold belt of Jurassic and Cretaceous strata borders the crystalline belt on the north.

The major geologic feature of southern Mexico and Central America is an extensive accumulation of Tertiary volcanic rocks which masks much of the underlying older rocks mentioned above. All the Isthmus of Costa Rica and Panama is made up of an igneous complex, mostly Tertiary, or of sediments derived from the volcanics.

Antillean Region

The Greater Antilles, composed of Cuba, Hispaniola, Puerto Rico, and the Virgin Islands, have a late Mesozoic and Cenozoic history. Thick limestones made up a northern facies in Jurassic and Cretaceous times and a volcanic assemblage a southern facies. Folding, thrusting, and intrusions followed. Tertiary time saw extensive flooding and reuplift of the islands but not much deformation of the strata.

The Lesser Antilles or Caribbees are a Cenozoic volcanic arc developed on the oceanic crust.

Precambrian (Plate 1)

Absolute age determinations on Precambrian rocks are now sufficiently numerous so that divisions of different ages are becoming defined. The ages denote the time of origin of the mineral of which the analysis was made, and this denotes the time of an igneous intrusion or of an episode of metamorphism. In other words, the ages appear to indicate belts of orogeny. They define a continent made up of a rather small central region of greatest age, and belts on the northwest and southeast of progressively younger age. The strange aspect of the belts older than 800 million years is that they project out to the Pacific Ocean basin, as if the continent at this time continued to the southwest farther than its present boundary. The Beltian basin or geosyncline, about 800 million years old, lies unconformably across the older belts, and apparently, for the first time marked a direction subparallel with the existing margin. Subsequently, all orogenesis occurred in belts conformable to the present margin.

Cambrian (Plate 2)

Cambrian seas and sediments defined major tectonic divisions of the continent which lasted until the end of the Paleozoic era. The Canadian Shield of Precambrian rocks formed the central and northeastern part of the continent, and it probably was a vast region of low relief. By way of an extension to the southwest, the Transcontinental Arch, the United States was divided into western and eastern seaways, and a symmetrial arrangement of shelves, miogeosynclines, and eugeosynclines resulted. No Cambrian strata are known along the western margin or the eastern margin south of Maine, and the conditions in these regions in Cambrian times are not well known. The hypothesis that North America was at this time part of a much larger continent, which cracked and spread apart, seems to help most in understanding the paleotectonic elements. Southern Europe, Africa, and South America are postulated by some to have lain close together, and hence, it is suggested that the Atlantic Ocean and Gulf of Mexico, with associated coastal plains or continental shelves, did not exist at this time.

Ordovician (Plate 3)

The broad Williston basin became well-defined during the Ordovician and a narrow basin of thick carbonate sediments formed in Oklahoma, and extended to the shallow Colorado sag nearly across the Transcontinental Arch. Extensive regions of the Canadian Shield were invaded by epeiric seas. The margins of the continent are still problematical. Flat-lying, unmetamorphosed strata in northern Florida south of the eastern orogenic belt seem to require continental connections where now is the Atlantic Ocean. The Taconic orogeny of folding and thrusting occurred in eastern New York, Vermont, and southeastern Quebec.

Silurian (Plate 4)

The Transcontinental Arch became very wide and well-defined. The Michigan Basin, in which extensive deposits of salines accumulated, and one in Pennsylvania and West Virginia took lasting form. The Ozark dome and Texas arch became prominent.

Devonian (Plate 5)

During Devonian time, the Transcontinental Arch rose, but was only gently emergent; and strata previously deposited across its site were removed by the close of the period, except for the Colorado sag where 100 to 200 feet of beds remained. In Canada, the great arch bifurcated into a broad arch west of Hudson Bay and another one east of Hudson Bay, but this condition probably did not arise until the close of the Devonian. During the Devonian the arches were at least partly submergent, because Manitoba fossil faunas are very similar to those of Michigan and the Hudson Bay region.

Transverse arches also developed. The Ellis–Ozark uplift extended from Kansas around the south end of the Illinois–Indiana–Kentucky basin to the Nashville dome, and thence northward to the Cincinnati dome. The Illinois–Indiana–Kentucky basin, with the intervening Kankakee arch or area of much less subsidence came into prominence for the first time.

A basin of subsidence centered in Pennsylvania during the Devonian, and sediments were supplied from the eastern Taconic orogenic belt which was being elevated adjacent to the basin and undergoing unrest premonitory to the Acadian orogeny. The basin sank mostly in late Devonian time, and its dominantly clastic and subaerial sediments coarsen toward the east. The western and northern marine facies constitute the classic Devonian section of the continent.

A Devonian trough extended northward through New England, the Maritime Provinces, and Newfoundland, in which much volcanic material was deposited along with various clastic sediments. In New England sedimentation was mostly east of the main Taconic belt, but in Quebec it occurred directly on the eroded Taconic structures. The entire region, beginning perhaps in mid-Devonian time in places, gradually became involved in the great Acadian orogeny. The strata were folded, intruded, metamorphosed, and thrust-faulted to form a complex of dominantly crystalline rock. The Acadian belt extended southward through the Crystalline Piedmont of the eastern United States, where numerous large batholiths were emplaced and considerable metamorphism occurred.

At the close of Devonian time a belt of orogeny, the Antler, formed in the central part of the western geosyncline. The belt continued active by way of folding and thrusting through Pennsylvanian, Permian, and Mesozoic time with a number of phases of orogeny fairly accurately documented. It effectively separated the miogeosyncline on the east from the eugeosyncline on the west.

Devonian and Silurian strata have been identified on the Pacific margin of the continent in the Klamath Mountains, and therefore it is concluded that the continental margin was then about where it is now.

Mississippian (Plate 6)

Mississippian seas were widespread and in the Rocky Mountain region a small basin subsided 10,000 feet along the Idaho–Montana boundary. A long eastward-extending basin sank through central Montana, and is known as the Big Snowy. A broad eugeosyncline of poorly known limits extended through northern California, southern Oregon, and northwestern Nevada west of the Antler orogenic belt. The amount of subsidence is unknown. The Antler orogenic belt in central Nevada was marked by major thrusting and complex folding.

The Transcontinental Arch sagged gently through its central area and was covered, but by the close of Early Pennsylvanian time it had risen enough to have suffered erosion, and the Precambrian was again exposed. The Texas arch was covered in central Texas and the Ozark–Nashville arch was severed from the Transcontinental Arch.

In latest Mississippian or earliest Pennsylvanian time, a deep and probably large basin sank rapidly in eastern Texas, southern Oklahoma, and western Louisiana, and received about 17,000 feet of clastic sediments.

The La Salle anticlinal belt first began to rise at the close of the Mississippian and continued to grow during the Pennsylvanian. It split the Illinois–Indiana–Kentucky basin in two parts.

The Ozark uplift developed into a broad, continuous arch with the Nashville and Cincinnati arches, and the northern arms of the Cincinnati arch, the Kankakee and Findlay arches, became well established. Gentle erosion probably occurred throughout this system of arches.

Subsidence continued in the Appalachian trough area, and a maximum of 4000 feet of sandstones, shales, and limestones accumulated.

In the Maritime Provinces and Newfoundland, a basin sank within the older Taconic and Acadian orogenic belts, and received about 5000 feet of clastic sediments, presumably from a rising orogenic belt on the east.

Pennsylvanian (Plate 7)

The south-central part of the continent was one of considerable and widespread unrest in Early Pennsylvanian time, and a number of ranges and basins were formed. The Wichita Mountain system of Oklahoma and northern Texas was uplifted together with the Ancestral Rockies of New Mexico and Colorado. The Pecos and Diablo ranges in west Texas appeared. The long, narrow Nemaha Range rose sharply, and at the same time basins on the east sank. The previously formed La Salle anticlinal belt was mostly buried.

The trough of the deep basin in eastern Texas of latest Mississippian and earliest Pennsylvanian time shifted northward to central Arkansas, and over 10,000 feet of sediments accumulated there.

The Arkansas basin was probably continuous with one in the southern Appalachians, where 10,000 feet of sediments, mostly clastic, accumulated. Such a thick and clastic deposit undoubtedly means vigorous uplift immediately on the southeast and south. The area of deposition in the southern Appalachians in Early Pennsylvanian time shifted to the central Appalachians in Late Pennsylvanian time, and somewhat more that 3000 feet of coal-bearing strata were deposited there. Although deposition had proceeded at variable rates here and there during the Paleozoic in the southern and central Appalachians, which lay to the west of the Taconic orogenic belt, it is generally stated that more than 30,000 feet of sediments had accumulated. In Late Pennsylvanian time or possibly in Early Permian time, the thick succession of strata from Oklahoma and Arkansas to Pennsylvania suffered folding and thrusting toward the continental interior, and the Ouachita Mountains and classical Appalachian Mountains (Valley and Ridge Province) were brought into being.

The Marathon orogeny of west Texas occurred in Late Pennsylvanian time and several thrust sheets moved northward toward the shelf. The Arbuckle Mountain system was formed by considerable folding and thrusting of the sediments of the Ardmore basin, and the structures were appressed tightly against the early Wichita Range.

In New England, the Maritime Provinces, and Newfoundland, subsidence followed somewhat the same pattern as that of the Mississippian, and coarse red Pennsylvanian clastics rest there on the Taconic and Acadian complex, and also in places in angular unconformity on the Mississippian.

Extensive subsidence occurred during the Pennsylvanian in the Cordilleran geosyncline with the deposition of more sand than in any time since the Cambrian. A local basin in west-central Utah subsided greatly and was filled in one place with about 25,000 feet of beds. The Antler orogenic belt dominated the sedimentary conditions east of it, and coarse clastics were spread there. Allochthonous masses were translated 25 to 75 miles eastward.

The volcanic orogenic archipelago persisted along the west margin of the continent, and was the source of volcanic contributions to the sediments of the adjacent seas, and the cause of unconformities and low-grade metamorphism in the deposits.

By Late Pennsylvanian time the Transcontinental Arch was almost entirely overlapped and buried, and the Early Pennsylvanian uplifts of Kansas, Oklahoma, and Texas were covered. Only the Ancestral Rockies in Utah, Colorado, and New Mexico remained as islands above the accumulating sediments.

Permian (Plate 8)

An eugeosyncline of deep and broad proportions developed in Permian time along the Pacific and was filled largely with volcanic materials. The Permian was a time of most extensive volcanism, and the site

of maximum subsidence and fill later became the locale of the great Nevadan batholiths.

Orogeny continued in central Nevada, and a small deep trough in western Utah filled with sandstone, shale, and limestone. Extensive shelf seas stretched eastward and southward.

The Colorado and Uncompahgre ranges of the Ancestral Rockies remained as islands in the surrounding seas.

The previously compressed Marathons were elevated epeirogenically, and in front of them several basins sank to considerable depth. The platforms of little subsidence between were the sites of the previous Pecos and Diablo ranges. The Anadarko basin also subsided appreciably.

The Carboniferous basins and adjacent areas of New England were intensely deformed either in Late Pennsylvanian or Permian time, and in places intruded by granitic batholiths. The deformation is not definitely dated, but presumably it occurred after the beds of the Permian basin of Pennsylvania and West Virginia had been deposited. It seems probable, also, that folding in Pennsylvania and West Virginia occurred at this time.

The crustal movements and spread of seas in Late Pennsylvanian and Permian time profoundly altered the geologic outcrop pattern of the continent. The greatest change comes from the extensive overlap of the pre-middle Pennsylvanian structures by the Upper Pennsylvanian and Permian sediments. All the Transcontinental Arch southwest of Wisconsin was buried, the structures of Kansas and parts of the Ozarks dome, the Wichita and Arbuckle mountain systems, and the ranges of west Texas vanished beneath the deposits. Only the Colorado and Uncompahgre ranges of the Ancestral Rockies remained visible, not because of renewed uplift, but because of considerable relief inherent from their original development.

Triassic (Plate 9)

Eugeosynclinal conditions continued in the west with extensive volcanic accumulations. Crustal unrest continued in central and western Nevada. In northern Utah a basin subsided and collected 8000 feet of carbonates and clastics. Eastward the Triassic deposits are largely continental and red. An emergent corridor connected the Canadian Shield with northern Mexico and southern Arizona.

The Colorado and Uncompahgre ranges still stood as islands in the surrounding deposits.

The Marathon–Ouachita orogenic belt of earlier development was still mountainous and had a broad piedmont generally free of deposits. The mountainous belt may have risen gently as its rocks were eroded and carried away, but orogeny there had ended.

Within the metamorphic and igneous core of the Paleozoic orogenic belts of the Atlantic margin, a zone of high-angle faults dropped basins and raised blocks of mountainous proportions. Volcanism was a prominent accompaniment of the faulting. The basins were the site of accumulation of thick, red, clastic sediments which were mostly derived from the uplifted, adjacent blocks. The basins are narrow and long, and because of their fault origin, their size was probably not much larger originally than now. The faulting and igneous activity ran its course in Late Triassic time, and the orogeny is known as the Palisades.

Jurassic (Plate 10)

The Cordilleran geanticline developed in Early Jurassic time and separated a western trough effectively from an eastern. The western again was one of extreme subsidence, and about 30,000 feet of volcanics, black shale, and other sediments accumulated in it. Central Nevada continued to experience orogeny, and thrusting of large proportions occurred. Late Jurassic was also a time of considerable batholithic intrusions in central and northern California and possibly western Nevada.

The eastern trough was generally the site of marine transgression and deposition, but the Jurassic deposits are less extensive than the earlier Permian, Triassic, and the later Upper Jurassic and Cretaceous. The Jurassic overlap on the Paleozoic strata of Montana, Alberta, and Saskatchewan, particularly on the Mississippian, is striking. The Mexican geosyncline began to form. It was separated on the north by a peninsula, the Coahuila, from the seas of the Gulf of Mexico. The wide basin of the

Gulf of Mexico had come into existence. Much salt was precipitated in an evaporite sequence in the Mexican geosyncline as well as along the northern part of the Gulf in Louisiana and Texas.

The interior of the continent was extensively emergent.

Early Cretaceous (Plate 11)

The Cordilleran geanticline widened and stretched from British Columbia to Mexico, and from eastern California to central Utah. A broad branch extended across Arizona and central New Mexico into Texas. Further deformation is noted in northwestern Nevada.

Basins sank greatly on the west in California, Oregon, and Washington. Volcanism continued there from previous times. The geanticline was flanked on the east by a trough of sedimentation from Alberta to northern Utah into which clastic sediments were shed. The Ancestral Rockies were buried save for a small island in central Colorado. The Mexican geosyncline enlarged and sank over 15,000 feet. It received considerable volcanic material from the west. The seas spread over the Coahuila peninsula to make it a platform, and were more extensive now over the southern and western part of the country than at any previous Mesozoic time.

The Rocky Mountain sea merged with the Gulf of Mexico, and the Gulf Coastal Plain sediments accumulated to an appreciable extent. Only the northern part of Florida was emergent. It was otherwise a platform and with the Bahama platform made up a large region of carbonate deposition and slow subsidence. It bordered on the south with a volcanic belt in Cuba where a carbonate facies on the north grades into a volcanic facies on the south.

Late Cretaceous (Plate 12)

The Late Cretaceous was a time of widespread and intensive crustal unrest along the western margin of the continent. The climatic phase of the Nevadan orogeny occurred at the very beginning of Late Cretaceous time when most of the batholiths that characterize the belt were intruded. Narrow basins subsided to considerable depths on the west margin of the belt where again volcanoes contributed some material. The Nevadan belt of orogeny became part of the broad Cordilleran geanticline, along whose eastern margin strong uplift with thrusting occurred. Floods of coarse conglomerate were poured into an adjacent trough in Utah and western Wyoming, and in places thrust sheets overrode the clastics.

The Late Cretaceous seas and deposits were even more widespread over the Rocky Mountains and Great Plains states than those of the Early Cretaceous, and the deposits were much thicker. East of the deep trough in central Utah, only thin deposits had previously accumulated under shelf sea conditions. Now sediments in excess of 5000 feet thick collected over a wide area of the shelf.

The Mexican geosyncline had shrunk and changed decidedly from its Early Cretaceous form. A trough extending from southeastern Arizona into northern Mexico contains much coarse conglomerate and volcanic material. South of the Coahuila platform a deep east-west trough, the Parras, sank and received over 15,000 feet of sediments, mostly limestones and shales.

Florida sank progressively through Late Cretaceous time, tilting southward to a trough that centered in Cuba where some 10,000 feet of carbonaceous sediments accumulated. As the carbonates thin northward through Florida, they change into argillaceous and arenaceous facies. The Atlantic margin of the continent was widely invaded, and a wedge of sediments that thickens seaward was deposited. The sediments overlap the Lower Cretaceous strata in most places.

Cretaceous-Tertiary Transition (Plate 13)

During the latest Cretaceous (Montanan) and earliest Tertiary (Paleocene) the main structures of the Rocky Mountains of Canada and the United States came into existence, and Plate 13 has been prepared principally to show these features. The crustal unrest is known as the Laramide orogeny. The Cordilleran geanticline was broadly deformed with its eastern margin and the adjacent basin deposits of the Triassic, Jurassic, and Cretaceous folded and thrust-faulted. Major overthrust

sheets rode eastward from the Yukon to southern Utah, and repeated floods of coarse clastics occurred in this marginal belt. Several phases of deformation are documented in most places.

East of the thrust belt including the large region from Montana to southern New Mexico and generally in the shelf region of sedimentation anticlinal uplifts, mostly elliptical in ground plan and asymmetrical in cross section rose in latest Cretaceous and Paleocene time. They are 75 to 150 miles long and 20 to 50 miles wide. Where the uplift has been great enough to result in erosion exposing the Precambrian rocks, thrust-faulting has occurred on the steep margin. The elliptical uplifts compose the major mountain ranges of the region. Between are intermountain valleys where, particularly in Wyoming and Montana, considerable amounts of Early Tertiary continental-type sediments were caught.

The western or Pacific margin of the geanticline continued to shed sediments to the adjacent basins, and no strong disturbance is indicated. The San Andreas fault had probably come into existence and the west-lying block at this time was lodged several hundred miles to the south, but now had started to shift northwestward along the fault, as indicated by the arrows.

Major deformation of previously deposited Jurassic and Cretaceous sediments occurred in the Greater Antilles with northward thrusting in Cuba.

Early Tertiary (Plate 14)

The most conspicuous and probably most significant feature of Early Tertiary time in the western cordillera was magmatic activity, especially volcanic. As can be seen from the map that the Great Basin region of Nevada, western Utah, and central and southern Arizona, together with the vast region of western Mexico, was mostly covered with volcanic materials. Southern Idaho was also extensively covered. Significant although scattered fields occur in New Mexico, Colorado, and Montana. The central Cordillera of Canada developed a large field. Several hundred small stocks also were intruded in the Great Basin, southern Arizona, and northern Mexico. All this activity followed the Nevadan and Laramide orogenies and, in places at least, marked the beginning of block faulting and rifting that dominated the Late Tertiary activities.

A eugeosyncline formed in Oregon and Washington, which is made up of a very thick mass of sediments and volcanics. Deep but restricted basins between uplifts developed in central and southern California. The San Andreas fault was very active and the west block moved northward, but was still considerably south of its present position. The Atlantic and Gulf of Mexico continental margins continued to subside during the Tertiary, but only in one or two places, particularly the Mississippi embayment, did the Tertiary beds overreach the Upper Cretaceous deposits. The Cretaceous and Tertiary sediments form the present Gulf Coastal and Atlantic Coastal Plains.

As the Atlantic margin of the continent subsided in Late Cretaceous and Tertiary time, the Appalachian orogenic belt arched gently, and successive erosion surfaces record the epeirogenic uplift.

The Greater Antilles sank and appeared as a belt of islands around which Tertiary sediments accumulated. Florida and the Bahama platforms also continued to sink and to be built up by carbonate sediments.

Late Tertiary and Quaternary (Plate 15)

Volcanism continued prominent in the Late Tertiary with basalt fissure eruptions in Washington and Oregon building the Columbia River field. To the south in southern Oregon and Idaho another extensive basalt field formed chiefly from vent eruptions. The west margin of these two large basalt fields has been built especially high by additional volcanoes to form the Cascade Range. A row of majestic stratovolcanoes of Quaternary age dominates the Cascades and extends into southern British Columbia beyond the basalt fields. The Cascade volcanics are chiefly andesite.

Block faulting of major proportions spread from the Sierra Nevada of California to the Wasatch Mountains of Utah. It also extended through southern Arizona and southward along the west coast of Sonora, Mexico. An arm of the faulting extended northward through eastern Idaho, western Wyoming, and western Montana to the Rocky Mountain Trench of British Columbia. The block, trench, or rift faulting is believed to be of tensional origin and to penetrate deeply into the crust.

The San Andreas fault block moved northward to its present position

and drifted apart from the continent at the south end to form the Gulf of Baja California which is floored by oceanic crust. Deep but local basins sank in southern California.

The Colorado Plateau block was uplifted with associated subsidence on the south and west. Several laccolithic groups were intruded into the Plateau strata, and several volcanic piles accumulated around the southern and eastern margins. In central Wyoming certain blocks were depressed along normal faults, particularly the Laramide Sweetwater Range. The Great Plains came into existence by uplift progressively greater toward the west. The Laramide Rockies were also uplifted, starting a new erosion cycle.

The marginal areas of the Gulf Coastal Plain continued to subside greatly under a heavy load of deltaic sediments. An area in northwestern Florida became emergent. The Atlantic Coastal Plain south of Long Island gradually rose and the sea retreated, but north of Long Island submergence and overlap of the sea occurred. The submergence has also been effective in Quaternary time southward where emergence has occurred previously.

Broad arching in the Appalachian region continued.

The Canadian Shield had been depressed under the weight of the ice sheets but in post-glacial time has lifted progressively to the north. The tilting starts at the hinge line shown on Plate 15 and amounts to 700 feet along the northern shores of Lake Superior, and possibly 900 feet along the eastern side of James Bay at the south end of Hudson Bay.

4.

PRECAMBRIAN TECTONIC PROVINCES

DISTRIBUTION OF PRECAMBRIAN ROCKS

The continent of North America is made up in a broad way of a stable interior and surrounding belts of deformed, intruded, and metamorphosed rocks. The stable interior has been free of orogeny since a time in the late Precambrian, or approximately for the last billion years. Before that time, however, a number of intense and widespread orogenies occurred.

The Canadian Shield is the greatest expanse of Precambrian rock exposures. These same rocks are blanketed by Paleozoic, Mesozoic, and Cenozoic strata over most of western Canada and the United States; only in areas of local uplift or doming have the old rocks been exposed. The Crystalline Piedmont of the Atlantic margin of the continent contains much rock of Precambrian age, and the western Cordillera exposes the ancient rocks of several ages and complex relations in a number of places.

CANADIAN SHIELD

Physiography

The Canadian Shield is characterized by a vast expanse of Precambrian rock. Its upland surfaces are uniform in height over large areas and, although now dissected, represent an old age erosion surface as large as any in existence today. The extensive surface rises 1000 to 2000 feet above sea level north of the St. Lawrence River and Lake Superior. Around Hudson Bay, especially on the south and west, is a wide lowland that ranges from sea level to 500 feet in elevation. In northern Labrador along the coast just southeast of Ungava Bay, the surface rises to 5000 feet and is extensively dissected. Hudson Bay is a great modern epeiric sea; it is a marine invasion from the north due to gentle subsidence in the heart of the shield in pre-Pleistocene or early Pleistocene time. The ice caps imposed such a weight on the shield in and around Hudson Bay that the area sank over a thousand feet in addition to the previous subsidence, and then with the melting of the ice it has risen about 900 feet.

Post-Proterozoic History

Paleozoic strata lap upon the shield from the Canadian plains on the west, and from the southwest in Saskatchewan and Manitoba. In northern Minnesota the Precambrian rocks lie exposed and extend southward into Wisconsin and eastward into northern Michigan. Paleozoic rocks continue to overlap the Precambrian across southern Ontario and Quebec to the Frontenac axis, where the Precambrian extends southeastward and forms the Adirondack dome in New York. See the Geologic Map of North America. For the most part, the Paleozoic rocks that skirt

the shield are Devonian and Silurian, and are chemical deposits or fine clastics. Along the southern margin of Hudson Bay is a fairly large area of flat-lying Devonian, Silurian, and Ordovician sedimentary rocks, and from fossil studies it seems probable that the Manitoba, Hudson Bay, and Michigan Devonian deposits were once continuous (G. M. Ehlers, personal communication). The thickness of the Devonian and Silurian south of Hudson Bay is at least 1000 feet, but their extension northward under the Bay's waters is not known. It can easily be imagined that they are continuous to Coats and Mansel islands at the entrance of Hudson Bay and thence to the nearly horizontal Paleozoic strata of Southampton Island and the Arctic Archipelago. If continuous, one wonders if somewhere in that large area the beds are not thick and form a trough or basin, perhaps similar to the Michigan basin. In fact, basins and arches have been recognized in the far north, and are described in Chapter 40.

It has been thought until lately that the Canadian Shield was comparatively free of epeiric seas in the past; but now, by the discovery of a number of small erosional remnants of Paleozoic strata far within the crystalline rocks (W. Sinclair and J. Tuzo Wilson, personal communications), it is believed that large areas were blanketed by sediments. Perhaps very little escaped submergence. What seems more important is that no orogenic belts developed across it during all of post-Proterozoic time. The same is true with some exceptions of the stable region of the United States.

In the iron ore belt of central Labrador (the Redmond iron deposit) downfaulting of a trench occurred in early Late Cretaceous time, and in it various argillites and ferric concretionary deposits accumulated. The Redmond deposit is in a basin 1 mile long, 1000 feet wide, and 600 feet deep. Abundant plant fossils in certain beds serve to date the deposits and the faulting. The extent of the Cretaceous faulting is not known (R. A. Blais, 1959).

From simple map examination, it looks probable that Greenland was part of the Canadian Shield until Cretaceous time when, perhaps, a Cretaceous trough extended as far north as Disco Island. Greenland was further severed from the shield either by Tertiary downfaulting or by drifting apart. See Chapter 40.

Geologic Provinces

The Canadian Shield until recently has been difficult of access, and this with the extensive "bush" cover has made geologic exploration expensive of energy and slow. The advent of airplanes and aerial photos has hastened the work immensely, and a beginning has now been made in analyzing the composition of the great Precambrian shield. But the time has not yet arrived, according to M. E. Wilson, when the vast region can be broken down into divisions with confidence. He draws approximate boundaries between five provinces (see map, Fig. 4.1), namely, the Western or Churchill, the Ungava, the Arctic Island, the Greenland, and the St. Lawrence. The last is divided into subprovinces, the North-

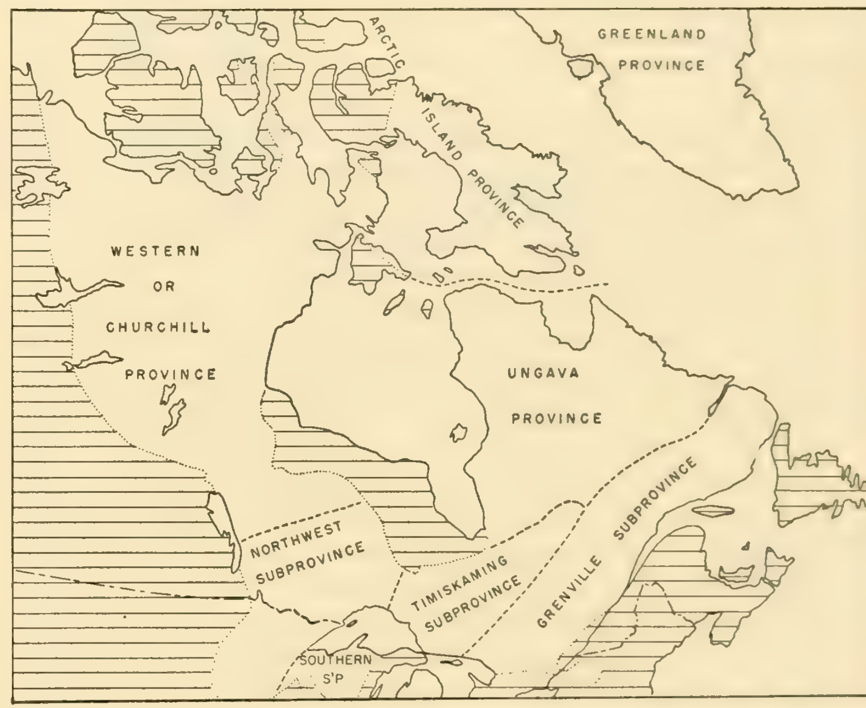


Fig. 4.1. Geologic provinces of the Canadian Shield best suited at present for individual formational names. After M. E. Wilson, 1958.

Era (10^9 Years)	Period-System	Major Sequence	Formation	Orogeny	Intrusive Rocks
Paleozoic	Cambrian				
			Unconformity		
(0.6 b.y.)			Hinckley sandstone		
			Fond du Lac sandstone		
			Unconformity		
(1.1 b.y.) Late Precambrian	Keweenawan	North Shore volcanic group	Undivided	Grenville	Duluth complex, sills at Duluth, Beaver Bay complex, Logan intrusives
			Puckwunge		
			Unconformity		Granite: St. Cloud Red, Rockville (?) granite at Granite Falls, Bellingham (?)
			Sioux quartzite (?)		
			Unconformity	Penokean	Gneiss: McGrath, Montevideo (?)
(1.7 b.y.)			Virginia argillite = Rove = Thomson		Tonalites: St. Cloud Gray, Warman, Hillman Freedhem, Montevideo
		Animikie group	Biwabik iron-formation = Gunflint		
Middle Precambrian	Huronian		Pokegama quartzite		
			Unconformity	Algoman	Granite: Gold Island, Giants Range, Sacred Heart, Fort Ridgely (?) Gneiss: Giants Range, Vermilion, Morton
(2.5 b.y.)	Timiskamian	Knife Lake group	Undivided		
(? b.y.)			Unconformity	Laurentian	Saganaga granite, Grassy Island tonalite (?)
			Soudan iron-formation		
Early Precambrian	Ontarian	Keewatin group	Ely greenstone		
		Coutchiching (?)	Undivided		
			Older rocks		

Fig. 4.2. Stratigraphic succession and geochronology of the Precambrian of Minnesota. Reproduced from Goldich *et al.*, 1961.

west, the Southern, the Timiskaming, and the Grenville. The provinces and the subprovinces thus defined represent natural divisions and the limits to which attempts should be made to correlate rock units. Wilson recommends separate names for formations, series, or intrusive bodies within each of these divisions at least for the present.

Geologists in recent culminating studies in the iron and copper region of Lake Superior recognize a threefold division of the rocks (Grout *et al.*, 1951, James, 1955, and Goldich *et al.*, 1961). The Precambrian of Minnesota is classified by Goldich *et al.* after many radioactivity age determinations, as shown in Fig. 4.2.

Previously, in 1934, a committee of the Royal Society of Canada on stratigraphical nomenclature had recommended that Precambrian time be divided into two eras, Archean and Proterozoic, and since then this classification has been used on most geologic maps issued by the Geological Survey of Canada. M. E. Wilson in 1958 contends that the dual classification is still the best and includes the Middle Precambrian rocks of Grout and James in the Proterozoic. In all provinces of the Canadian Shield a profound unconformity is known, by reference to which the rocks can be divided into two great groups (M. E. Wilson, 1958). The standard of reference, for instance, is the rock succession on Lake Timiskaming where the Huronian (Cobalt series) rests "with great unconformity" on granite—the Laurentian.

The Archean rocks consist of clastic sediments and various volcanic rocks conformably interbedded, and even though extremely old they are so little affected by metamorphism in places that original sedimentary structures are clearly visible. In many other places they are metamorphosed to various degrees. According to Pettijohn (1943) one may study the bedding in certain argillites in the finest detail, and the associated volcanics show pillow structures, amygdules, spherulitic structures, the same as in lavas of much later geologic time. Metamorphism is mainly of the low-grade variety, and orogeny has left the very ancient rocks of many areas untouched. Recognizing the near-absence of metamorphism in places, however, it must also be understood that enormous volumes of intrusive igneous rocks occur, and estimates have been made that these intrusive rocks constitute as much as 80% of the shield. The great bulk of these are granites of various types, with relatively small but important amounts of basic rocks such as gabbro, norite, and peridotite. Needless to say, much metamorphism has occurred and gneisses and schists (migmatites) are extensively developed.

> The Archean sediments of the southern Canadian Shield are mainly graywacke. Much conglomerate, a little slate, and still less iron-bearing formation are also present. Excessive thickness, especially of the conglomerates, abundance of graded bedding, rarity of cross-bedding and absence of ripple mark, the graywacke nature of the arenaceous beds, the absence of true quartzites and limestones, and the scarcity of normal argillaceous sediments, and the association with greenstones and tuffs are all the earmarks of a geosynclinal facies of sedimentation (Pettijohn, 1943).

In particular, these types characterize the eugeosyncline, and since they are repeated in later Precambrian rock series, it is little wonder that confusion in correlations has resulted.

In eastern Ontario and adjacent parts of Quebec the oldest rocks are sedimentary gneisses associated with great thicknesses of crystalline limestone and a little basic metavolcanics. These rocks are termed the Grenville series. They appear to have been originally shales, sandstones, limestones, and some lavas, but owing to the intense metamorphism, they are now biotite schists and sillimanite-garnet gneisses, vitreous quartzite, and crystalline limestones.

In southern Ontario, particularly in Hastings County a younger series, the Hastings, overlies the Grenville with erosional unconformity but, apparently with little structural discordance. The series consists of gray, blue-weathering limestone interstratified with argillite, except near the base where beds of congolmerate interstratified with argillite, buff-weathering dolomite, graywacke, and mica schist occur. Both Grenville and Hastings rocks are intruded by a group of gabbros, anorthosites, pyroxene diorites, and pyroxene syenites. Later still are dikes, sills, and batholiths of granite and syenite, and their gneissic equivalents.

The Grenville subprovince is believed to be separated from the Timiskaming subprovince by a great fault called the "Lake Mistassini–Lake Huron fault" by M. E. Wilson (1956) and the Grenville front or fault zone" on the Tectonic Map of Canada (1950). The fault marks a zone of considerable disturbance, and in the Lake Mistassini area it seems evident that the Grenville rocks have been thrust over those of the Timiskaming subprovince. The theoretical fault lies under lakes and glacial deposits for most of its length, and considerable controversy centers about it.

For further discussion of the many rock units already described over the vast Canadian Shield read M. E. Wilson 1956 and 1958, and Harrison, 1957. A recent symposium publication, "The Grenville Problem," published by the University of Toronto Press, presents a fascinating picture of the many problems involved.

Tectonic Provinces

With the advent of physiochemical age determinations (about 1931) much new light has been shed on the relative ages of rocks in the Canadian Shield. The ages are actually for minerals occurring in igneous rocks or in reconstituted rocks, metamorphosed during an orogeny; the original age of the graywacke, shale or lava is not determined but rather the age of the orogeny. Therefore, with the absolute age determinations has come an increased attention to orogenic belts, and certain geologists have postulated a division of the Canadian Shield into tectonic provinces or orogenic belts, in place of the "geological provinces." See Fig. 4.3.

The oldest orogeny in Minnesota is called the Laurentian by Goldich *et al.* (1961), but this he regards as an early phase of folding to the

Fig. 4.3. Precambrian orogenic belts of North America defined by isotope ages.

greater Algoman orogeny (see Fig. 4.2). The latter occurred about 2500 m.y. ago, although the very ancient dates range from 2200 to 2600 m.y. The name Algoman is here used for the belt of ancient dates through the southern part of the Canadian Shield. It has been variously called the Keewatin and Superior by other writers.

The Algoman and the Slave (also called Yellowknife) provinces are the oldest known and possibly parts of the original nucleus of the continent. They have a high ratio of lavas to sediments which are of the graywacke facies, presumably deposited in geosynclinal basins. The Churchill province is considered an orogenic belt by which the two nuclei were welded together (J. Tuzo Wilson, 1949, 1954). See also Farquhar and Russell (1957) and Lowdon (1960).

A belt of Huronian rocks extending from Minnesota through Wisconsin into Michigan and lying south of the main Algoman belt has ages of about 1700 m.y. Goldich *et al.* (1961) call it the Penokean orogenic belt, and the name has been applied in this book to adjacent regions on the southwest in the United States and on the northwest in Canada.

The Grenville subprovince of M. E. Wilson approximately is postulated as an orogenic belt about 1000 m.y. old. Its deformed front borders directly on the Algoman province. Southeast of the Grenville belt are the Taconic and Acadian orogenic belts, about 400 and 300 m.y. old, respectively.

Eighty-three isotopic age analyses on biotite, K-feldspar, and whole-rock samples from forty-five localities, using both K–Ar and Rb–Sr methods have been made on igneous rocks and a few metasediments in the Sudbury–Blind River area of the Grenville belt by Fairbairn *et al.* (1960).

> The numbers obtained, forming an almost continuous age spectrum from 1.0 b.y. to 2.2 b.y., are correlative with widespread and repeated diastrophism in the region. Whole-rock analyses of igneous material, where available, show higher ages than coexisting minerals in most examples, and there is reason to believe that these are close approximations to the true age. There is considerable evidence by both K–Ar and Rb–Sr methods, of orogenic events at approximately 1.0 b.y., 1.2 b.y., and 1.6 b.y.
>
> The oldest igneous rock found thus far is the Copper Cliff "rhyolite" (2200 m.y.), which intrudes the basal section of a thick series of conformable metasediments and volcanics southeast of Sudbury. At Quirke Lake granite in the basement, uncomformably beneath U-bearing pebble beds, is 2050 m.y. old. As the time of uranium mineralization in these Huronian sediments is placed at 1700 m.y., and gabbro which intrudes them may possibly be older than 1800 m.y., their deposition must have been in the age bracket 1800–2050 m.y.

ARCTIC STABLE REGION

South of the orogenic belt of northern Greenland and Ellesmere Island and north of the Precambrian Canadian Shield is a stable region composed of a Precambrian crystalline basement with a veneer of nearly horizontal Paleozoic sedimentary rocks. It includes most of the Arctic islands, and the shallow sea-covered areas between. See the *Geologic Map of North America* or the *Geologic Map of Canada*. The Precambrian rocks of the shield extend northward into Baffin and Devon islands, and exposed extensively in Melville and Boothia peninsulas, but the Paleozoic blanket indicates that much, if not all, of the Arctic islands region (also called Arctic Archipelago) and the northern part of the Canadian Shield were submerged at times during the Paleozoic. The part south of the fold belt (Chapter 35) has suffered only gentle vertical movements since the Proterozoic, and is therefore part of the great stable interior of the continent. The Precambrian crystalline rocks extend southward into the United States under a veneer of Paleozoic sedimentary rocks commonly called the Central Stable Region. It seems appropriate, therefore, to speak of the similar northern geologic province as the Arctic Stable Region.

PRECAMBRIAN PROVINCES OF THE UNITED STATES

Isotope Age Determinations

Recent age determinations fall into a pattern that marks successive orogenic belts in the central, southern, and western states of the United States, and these are shown in Fig. 4.3. The ages pertain to rocks generally called Archean or basement complex. In Arizona, Utah, Idaho, and Montana, younger and much less metamorphosed strata rest unconform-

ably on the crystalline basement, and are variously called Algonkian, Proterozoic, Beltian, or Upper Precambrian. These are shown on the map by the dotted lines. Extending southwestward from the western part of Lake Superior is another belt of late Precambrian rocks, namely the Keweenawan Series with its included large gabbro sills. Beneath the Paleozoic and Mesozoic sedimentary cover of Texas and southeastern New Mexico still other young Precambrian sediments, volcanics, and gabbro sheets have been recognized, resting on an older granitic terrane.

Algoman Orogenic Belt

The ages thus far published for north-central Wyoming and south-central Montana are very old (2500 to 2760 m.y.) and stand apart from other ages in the Rocky Mountains (Aldrich *et al.*, 1957; Gast and Long, 1957; Hayden and Wehrenberg, 1959). An absolute age determination in southeastern Manitoba between Winnipeg River and Johnston Lake indicates that a plutonic and metamorphic cycle occurred 2650 ± 100 m.y. ago (Eckelmann and Gast, 1957). These ages are 400 to 500 m.y. older than those recorded for the "Superior" Province in Canada, but even so are much closer to it than to those of the adjacent younger orogenic belt, and hence are regarded related.

Penokean Orogenic Belt

A number of isotope ages to date seem to establish an orogenic belt of intermediate age between the very old Algoman and the younger Mazatzal. These are in the range of 1600 to 1750 m.y. See Fig. 4.3. The belt contains a mixture of the old dates and the younger, and this is taken to mean that the younger orogeny was superposed on the older. The analyses are so few to date that the northern limit of the belt is poorly defined, and not much reliance for tectonic interpretive purposes can yet be placed on the distribution. The southern limit is somewhat better defined, with none of the older dates in the general field of the 1250 to 1450 m.y. dates.

On the basis of the geology of the rocks of southwestern Montana the two ages are understandable. Perhaps even more ages within the belt will be recognized. A brief description of the recognized units is as follows:

(1) The oldest units underlie the Cherry Creek Group and include rocks of the Pony Group as well as other pre-Cherry Creek rocks which probably are not time equivalents of the Pony. Main types present are biotite-garnet gneiss, granite gneiss, injection gneiss, and amphibole gneiss. (2) The Cherry Creek Group consists of metasediments including marble, quartzite, micaceous schists, sillimanite schist, banded ironstones with intercalated layers of amphibole gneiss, and amphibolite representing metamorphosed mafic sills and flows. (3) A number of post-Cherry Creek intrusives, all of which show varying degrees of metamorphism, include, among others, the Dillon granite gneiss, widespread in Beaverhead and Madison counties, the granite of the Jardine district, and the Pinto metadiorite in the Little Belt Mountains. (4) Widely distributed bodies of unmetamorphosed peridotite and associated ultramafic rocks have as their largest representatives the Stillwater Complex. (5) Post-Stillwater intrusives are represented mainly by the granite of the Beartooth Range. (6) Numerous and widespread diabase dikes that cut all these older units but do not extend into Beltian rocks (Heinrich, 1953).

The crystalline basement of the Beartooth Range from what is known consists of schists and gneisses, possibly the Cherry Creek. On the northeast is the Stillwater ultramafic complex which has been intruded into a series of dense gray hornfels, an iron formation, and light-colored quartzites. It may be part of the Cherry Creek group. A light gray gneissoid biotite granite cuts the ultramafic complex. At Cook City two granites are recognized (Parsons and Bryden, 1952).

The roof of a granitic batholith is exposed in the Teton Range of western Wyoming. The deep canyons that dissect the range show gigantic zenoliths and an irregular roof of gneiss and schist.

Mazatzal Orogenic Belt

Distribution of Dates. A good scatter of age determinations has been made in the Rockies from the Black Hills to Arizona and southern Nevada and defines a belt of rather consistent age between 1300 and 1400 m.y. old. A low age is given for the Front Range of central Colorado of 1100 m.y., a high age for the Black Hills of South Dakota of 1600 m.y., and a high age of 1590 m.y. for the Central Wasatch Mountains in Utah. Other than these three, ten other ages fall fairly close to 1350 m.y.

No orogenic belt or province in the Canadian Shield has yielded such ages. The 1350-m.y.-old belt of the western United States appears to

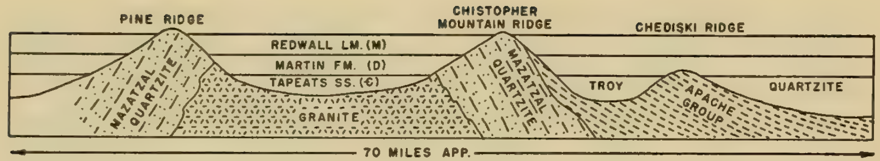


Fig. 4.4. Restored section across the northern part of Mazatzal Land. After Huddle and Dobrovolny, 1950.

project into the Grenville belt or wedge out to the northeast in the Great Lakes region.

Arizona. The 1350-m.y. orogenic belt is here called the Mazatzal from relations in central Arizona (Fig. 4.4). A correlation of the Precambrian rocks of Arizona by Anderson (1951) is given in Table 4.1, and in it will be seen that the Mazatzal quartzite is regarded as the youngest of a group of old rock units, mostly schists. E. D. Wilson (1939) showed that

Table 4.1. Correlation of Precambrian Rocks of Arizona (C. A. Anderson, 1951)

	Grand Canyon (Noble and Hunter 1917; Darton, 1925)		Bradshaw Mtns. (Lindgren, 1926)	Mazatzal Mtns. (E. D. Wilson, 1939)		Globe (Ransome, 1903)
Younger Precambrian	Grand Canyon Series	Chuar group Unkar group		Apache group		Apache group
Unconformity						
Orogeny—Intrusion of granitoid magmas						
Older Precambrian	Vishnu schist		Yavapai schist	Yavapai group	Mazatzal quartzite Maverick shale Deadman quartzite Alder series Red Rock rhyolite Yaeger greenstone	Pinal schist

granite, and thus dated the orogeny and intrusions as post-Mazatzal. the Mazatzal quartzite was folded and faulted prior to the intrusion of He named the orogeny the Mazatzal revolution, and this event now seems to be dated by the new isotope age determinations, and therefore is applied to the entire belt up through Colorado, Wyoming, and South Dakota.

It should be noted that the Vishnu schist is 25,000 feet thick where exposed in the Grand Canyon of the Colorado, and was originally fine-grained argillaceous sandstones and sandy shales. A sequence of basaltic lavas and tuffs is now represented by amphibolites in which relict pillow and anygdaloidal structures prove the volcanic character. The Vishnu schist is intruded by plutonic rocks that range from quartz diorite to granite. In fact, granite is more widespread in outcrop in Arizona than the host rocks, and therefore the Mazatzal orogeny must be considered, there at least, to be identified with great batholithic intrusions of fairly acidic rock.

Colorado. The largest exposure of basement crystalline rocks in the Rockies is in the core of the Front Range of Colorado. It consists essentially of granite, schist, and gneiss (Lovering and Goddard, 1950).

The oldest rocks in the Front Range are the schists and gneisses of the Idaho Springs formation, which are highly metamorphosed sedimentary rocks of early pre-Cambrian age. The thickness is approximately 20,000 feet. The hornblende schist and gneiss of the Swandyke hornblende gneiss is overlain by a series of quartzites and quartz pebble conglomerates at least 14,000 feet thick. These formations are all cut by an extensive series of granite intrusives, the oldest of which is a quartz monzonite gneiss. It occurs chiefly in small stocks peripheral to granite batholiths or as a lit-par-lit injection of the older schists and gneisses. Gneissic granite, gneissic aplite, and gneissic diorite are found in abundant but small masses within the metamorphic terrain and are believed to be related to nearby granite batholiths of different ages.

The earliest of the batholithic granites is the Boulder Creek granite; it is common in stocks and small batholiths in the central part of the Front Range. Its dark-gray color and faintly banded appearance distinguish it from the pink coarse-grained Pikes Peak granite, which is somewhat younger and forms the extensive batholith of the southern part of the Front Range. The appearance and age relations of the Pikes Peak granite are the same as those of the Sherman granite exposed in the large batholith extending from the northern part of the Front Range well into Wyoming. Small batholiths and stocks of the younger fine-grained to medium-grained light pinkish-gray Silver Plume granite are

widely distributed, and locally have been given different names. Lead-uranium ratios indicate that the age of the Pikes Peak granite is approximately 1 billion years and that of the Silver Plume granite approximately 940 million years (Lovering and Goddard, 1950).

The lead-uranium ratio age determinations for the granites are younger than those yielded elsewhere by the potassium-argon and rubidium-strontium methods, and it seems probable that these will be recognized as too young and replaced by new age determinations.

Utah. The Precambrian rock succession in central Utah is shown in the correlation chart of Fig. 4.5. The Farmington Canyon complex is the basement rock and consists of gneisses, schists, and granulites, about 20,000 feet thick, once a stratified sequence of arkose, calcareous shale, impure dolomitic and tuffaceous beds, and very pure quartz sandstone. Metamorphism is of the lower amphibolite facies and therefore medium-grade (Larson, 1957; Bell, 1951). The metamorphism is dated as 1590 m.y. (Gast and Long, 1957).

Another sequence of beds, the Willow Creek and Harrison, seems to be of intermediate age, and it is not clear yet whether they were involved in the Mazatzal orogeny. The Farmington Canyon complex is overlain unconformably by the Big Cottonwood quartzite and argillite series and did not participate in the metamorphism of the older gneisses and schists.

The Big Cottonwood and Uinta series are generally correlated with the Belt series of western Montana which is very thick and widespread. These will be referred to under the next heading.

Beltian Orogenic Belt

A major trough or geosyncline of sediments and volcanic rocks of post-Mazatzal age, yet pre-Paleozoic age, extends north and south from the Mexican border through Arizona, Utah, Idaho, western Montana, eastern Washington, western Alberta, and eastern British Columbia to the Yukon, and possibly into Alaska. Its stratigraphy is complex, and much remains to be discovered and worked out. Two major divisions appear to stand out, namely, a lower one, the Beltian, and an upper one, which is typified by a thick and well-described succession in the western Purcell Range (Reesor, 1957) and in northeastern Washington (Park and Cannon, 1943). It has been referred to as the Upper Purcell by Reesor (1957) and also as the Lipalian series by Gussow (1957). In northern Utah, it may find representation in the Mineral Fork tillite and Mutual formation (Crittenden *et al.*, 1952).

Angular unconformities have been recognized in a number of places up and down the trough between the Beltian and Metaline sequences and between them and the overlying Cambrian. In central Arizona (Mazatzal Mountains) the Apache (Beltian) group is tilted, beveled, and overlain by the Cambrian. In the Grand Canyon of northern Arizona, the Grand Canyon series (Beltian) group is tilted, faulted, beveled, and overlain by the Cambrian. In north-central Utah 12,000 to 15,000 feet of the Big Cottonwood series (Beltian) and the Mutual formation are cut out beneath the basal Cambrian angular unconformity.

In western Montana and southeastern British Columbia Deiss (1935) believes the Beltian strata were strongly uplifted, tilted, mildly folded, and eroded before the Cambrian beds were laid down. In the Purcell Range Cambrian beds lie across various Purcell formations (Beltian) through a stratigraphic interval of 8000 feet, and although the discordance is generally slight, in one place it is 90 degrees (White, 1959). Large sills and dikes are present in this region and probably accompanied the orogeny. Campbell (1959) recognizes an unconfromity between Middle Cambrian and Beltian strata in northwestern Montana and northern Idaho in which up to 18,000 feet of Beltian is missing.

Stimulated by a paper by Weiss (1959) the writer has prepared a cross section from northeastern Washington across southern British-Columbia to Waterton, Alberta, showing postulated conditions at the beginning of Middle Cambrian time (Fig. 4.6). The Beltian correlatives would be the Deer Trail, Priest River, and Lower Purcell groups. The Upper Purcell group would include the Monk, Three Sisters, and Horsethief Creek association, the Huckleberry, Leola, Irene, and Purcell volcanics, and the basal Huckleberry, Shedroof, and Toby conglomerates. It may be seen that the Upper Purcell group rests unconformly on the Beltian and is introduced by a thick and widespread conglomerate. This unconformity is taken specifically to mark the orogeny of the Beltian

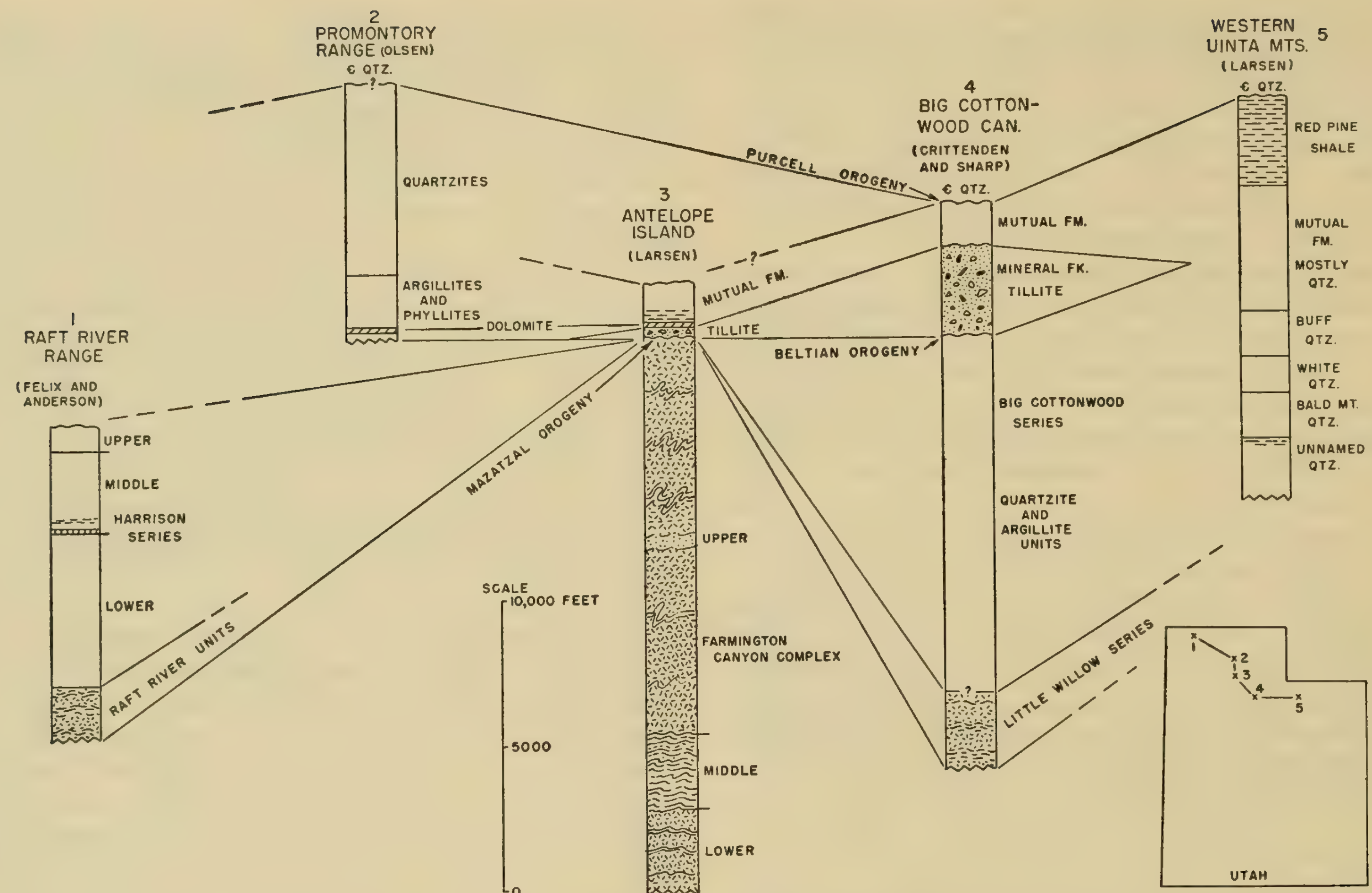


Fig. 4.5. Correlation chart of Precambrian formations in northern Utah. After Larson, 1957.

orogenic belt, and its extent is assumed to be approximately that of the Beltian trough. The conglomerates appear to have come from the west, and if so, the orogeny was most severe along the western margin of the trough.

The age of the Beltian orogeny cannot be accurataely fixed with existing data. A sample of illite from a shale in the Siyeh formation in Glacier National Park (Goldich *et al.*, 1959) yielded a date of 740 m.y. by the potassium-argon method and 780 m.y. by the strontium-rubidium method. Goldich *et al.* reason that this age is not a time of metamorphism but more probably marks the time of deposition. The Siyeh formation is near

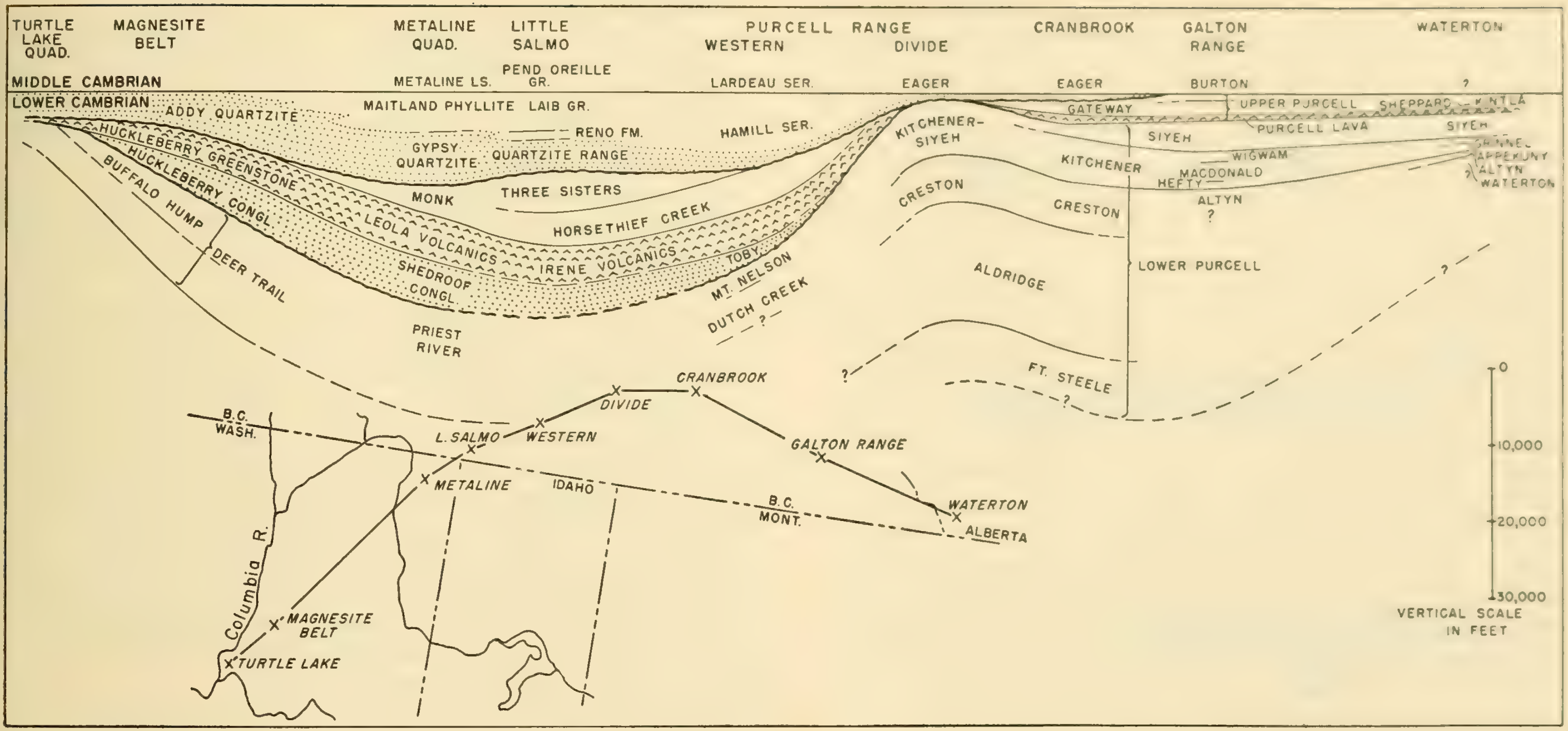


Fig. 4.6. Suggested correlation of Precambrian formations of southern British Columbia and northeastern Washington, after Reesor (1957) and Weiss (1959), restored to Middle Cambrian time.

the top of the Belt series, and the Beltian orogeny occurred soon after its deposition, so the date is about as good for the time of deposition as for the metamorphism, if any, or orogeny.

In conflict with the illite date we note that samples of uraninite in a vein system in the Coeur d'Alene district of Idaho that cuts folded meta-sedimentary rocks of the St. Regis formation of the Belt series have yielded a date of approximately 1190 m.y. (Eckelmann and Kulp, 1957). Although different laboratories have confirmed this date, Wehrenberg (personal communication) thinks there is still justification to question its validity in dating the age of the strata and their folding. The St. Regis is about three-quarters of the way up from the lowermost beds of the Belt exposed. From samples of galena in the same mine Farquhar and Cummings (1954) give the age as 1030 ± 290 m.y.

It is clear from Fig. 4.3 that the Belt sediments and correlatives lie in

a great elongate basin generally north-south and parallel to the Pacific margin of the continent, and that the basin is discordant with the older orogenic belts, across which it lies. This, if true, is of great significance. It suggests that following the Mazatzal orogeny that a major part of the western margin of the continent was removed, because the older belts of orogeny now extend at nearly right angles to the continental margin. It also suggests that in Beltian time the processes of sedimentation and orogeny first became established along and parallel to the present continental margin.

The discordant relation of the Beltian trough and orogenic belt to the older belts emphasizes the concern that must be attached to the uraninite date. It is almost as old as the Mazatzal orogeny, and presumably should be separated from it by considerable time.

Not only is the Beltian orogenic belt discordant with the Mazatzal orogenic belt, but also are the Antler and Shuswap belt and Nevadan belt which lie west and parallel with the Beltian (see Chapters 6 and 17). If the theory is held that the nucleus of the continent has been added to by successively younger orogenic belts, then some major change occurred to the southwest margin of the North American continent in Beltian or pre-Beltian time. Perhaps a major part of the southwest margin as it existed in pre-Beltian time is missing, but no plausible theory of translation or foundering has been thought of to restore the missing part. It is conceivable that a major change occurred in the constitution and assembly of the continents in the interval of time immediately preceding the Beltian.

Purcell Orogenic Belt

Following the Beltian orogeny in the southern British Columbia and northeastern Washington region a thick conglomerate was deposited, and then extensive volcanic rocks were spread all the way from the Columbia River in Washington to Waterton, Alberta. These were followed by sandstones and argillites, particularly in a main trough in the Purcell Range area. After this depositional and volcanic cycle another disturbance occurred in which, in the Purcell Divide area, the entire series was removed together with a considerable thickness of the underlying Belt series (Weiss, 1959). This unconformity attests the removal of a greater thickness of strata than the one at the base of the Shedroof–Toby conglomerates, according to Weiss. See Fig. 4.6.

The overlying Lower Cambrian quartzite appears to have been derived from the west, like the basal Huckleberry–Shedroof–Toby conglomerate, and, if so, indicates that the major axis of orogeny lay to the west. The zone from the Purcell Range to the front of the present Rockies was a broad geanticline across which the Early Cambrian seas failed to spread. The Middle Cambrian seas, however, probably transgressed much of the geanticlinal area (Campbell, 1959).

The orogeny of post-Monk and Three Sisters age, yet of pre-Early Cambrian age, will here be called the Purcell.

In dealing with Precambrian formations distant correlations are generally questionable, and this is especially so when assuming that the Mineral Fork tillite and Mutual strata of northern Utah are equivalent to the Upper Purcell group. If valid, however, an orogeny can be said to have occurred after the close of Mutual time and before the late Lower Cambrian sands were spread across the beveled edges of these formations as well as those of the Big Cottonwood series. It is not clear how discordant the tillite and Mutual are to the underlying Big Cottonwood strata because of limited exposures, but Crittenden *et al.* (1952) note that the tillite occupies broad smooth-bottomed basins scooped out of the upper part of the Big Cottonwood series.

Both the Beltian and Purcell orogenies may be combined in one angular unconformity in the Grand Canyon of the Colorado in northern Arizona. It is evident that information on the extent of the Beltian and Purcell orogenies in scanty and that the pronouncements of the preceding paragraphs are postulates of fairly tenuous nature.

Keweenawan Belt

The Keweenawan series of the Lake Superior region is the youngest of the Precambrian rocks there and is well known because of the great value of its copper mineralization. An imposing sill dated 1100 m.y. by Goldich (personal communication) and believed to be part of the Keweenawan series, crops out along the northwest shore of Lake Superior.

It is called the Duluth gabbro. Three divisions of the Keweenawan are recognized, namely, a lower clastic sequence 1400 feet thick, then a thick unit of basic amydaloidal lava flows interbedded with sandstones and conglomerates, and at the top a continental clastic sequence possibly reaching a thickness of 25,000 feet in the center of the basin of accumulation. The widespread extent of the flows and the paucity of ash suggest that the flows issued from a system of fissures rather than central vents. Associated with the flows and intruded into them are numerous dikes and sills, dominantly basic. The most prominent sill is the Duluth gabbro.

The thick upper Keweenawan clastics consist of red feldspathic shaly sandstones at the base and these grade upward into arkosic and quartzose sandstones. They accumulated as the basin foundered in response, presumably, to the extrusion of the large volume of volcanics. Highlands existed on both sides of the basin (Hamblin and Horner, 1961).

Several large faults break the Keweenawan series. The Douglas and Keweenawan are found on opposite sides of the synclinal or basin axis with thrusting away from the axis. See map, Fig. 4.3 and cross sections of Fig. 4.7. Vertical displacements up to 4 miles are indicated by the cross sections. The North Shore fault, postulated from physiographic data solely (principally from the straight shorelines) was not detected by gravity surveys, but the surveys do not rule out its existence. If it is a reality, it may be a normal fault and of later age than the reverse faults. The orogeny of post-Keweenawan time consisting of volcanism and faulting has been called the Killarnean and is dated at about 950 m.y. (Fairbairn *et al.*, 1960).

The sills and volcanic rocks are strongly reflected by positive gravity anomalies, and the deep basins of clastic rocks by negative anomalies. Thiel (1956) has recognized this fact and traced the Keweenawan series under the Paleozoic sedimentary rock cover by means of these strong anomalies southwestward to the Salina basin of Kansas. The positive feature has an average width of 30 miles and an amplitude of 100 miligals above the regional gravity value. For the greater part of its length it is flanked on both sides by gravity lows. The igneous rock masses are responsible for the gravity highs and the clastic-filled basins, the lows.

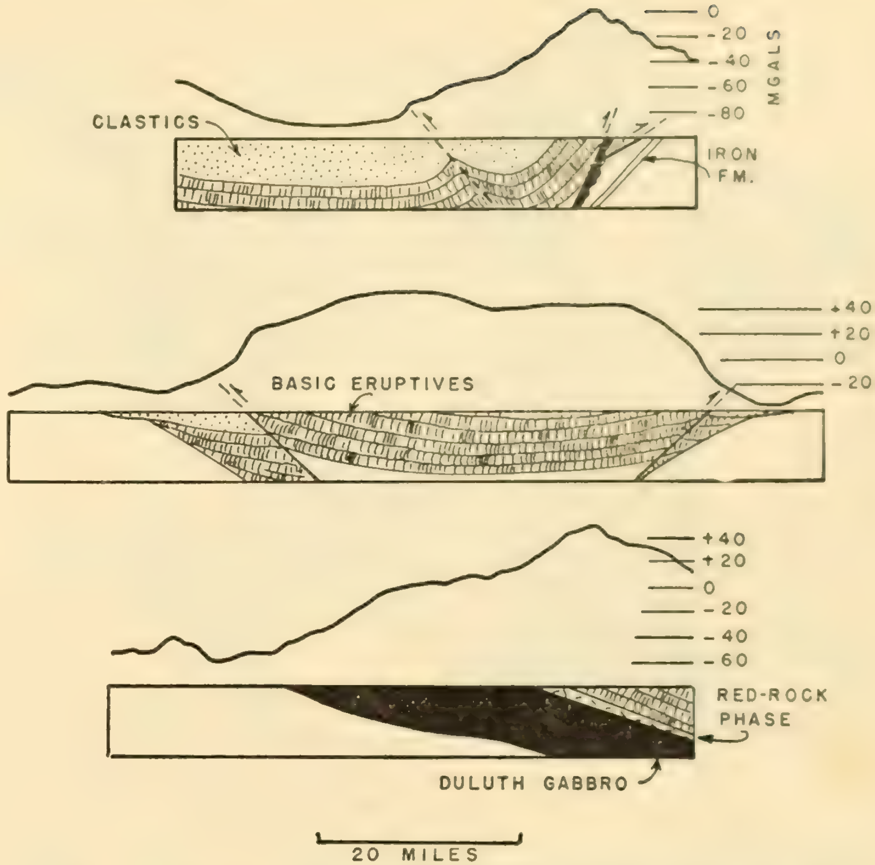


Fig. 4.7. Keweenawan orogenic belt. Sections in the Duluth area after Thiel, 1956.

The Keweenawan belt projects toward the volcanic and gabbroic terranes of Oklahoma and Texas, and perhaps these are part of the same tectono-igneous belt. No strong gravity anomalies are known between Kansas and Texas, but the grain of gravity contours (Lyons, 1950) is southwesterly, and thus the belt may be marked by sedimentary rocks and an absence of volcanic in this region.

The Precambrian rocks of the Wichita Mountains of Oklahoma represent the upper granitic part of a large gabbroic lopolith which

	CENTRAL TEXAS	NORTH TEXAS	TEXAS PANHANDLE	WICHITA MOUNTAINS—BURIED AMARILLO MOUNTAINS	ARBUCKLE MOUNTAINS	VAN HORN AREA	WEST MARGIN OF TEXAS CRATON; FRANKLIN MOUNTAINS; SOUTHEAST NEW MEXICO
						sedimentary rocks (Van Horn sandstone)	
			SWISHER GABBROIC TERRANE emplacement of gabbro (lopolith?); contact metamorphism of sedimentary rocks	WICHITA IGNEOUS PROVINCE gabbro-granite (670 m.y.) intrusion; contact metamorphism of sedimentary rocks (Meers quartzite)		local orogeny—cataclastic metamorphism; diorite intrusion	
			subsidence; sedimentary rocks (carbonate rocks and siltstones)	sedimentary rocks Meers quartzite)		sedimentary rocks (Allamoore and Hazel formations)	
LATE PRECAMBRIAN			PANHANDLE VOLCANIC TERRANE lavas, tuffs, shallow intrusives—mostly rhyolite		rhyolite intrusions? (East and West Timbered Hills porphyries)	rhyolite intrusions	rhyolite intrusions and extrusions
	FISHER METASEDIMENTARY TERRANE regional metamorphism of sedimentary rocks	RED RIVER MOBILE BELT regional metamorphism of sedimentary rocks; intrusion			synorogenic? granite intrusions	VAN HORN MOBILE BELT regional metamorphism (Carrizo Mountain group pre-rhyolite)	regional metamorphism of sedimentary rocks (Lanoria quartzite?)
	TEXAS CRATON granitic intrusions (about 1000 m.y.)	Texas craton to south	TEXAS CRATON granitic intrusions			Texas craton to north and northeast	Texas craton to east
MIDDLE PRECAMBRIAN	regional metamorphism and intrusion (Valley Spring gneiss, Packsaddle schist, older gneissic meta-igneous rocks.						

Fig. 4.8. Tentative correlation of Precambrian rocks and structural events in Texas, southern Oklahoma, and southeast New Mexico. Reproduced from Flawn, 1956.

Hamilton (1956c) thinks might correlate with the Duluth gabbro.

The lithologies and age relations recognized by Flawn (1956) of the Texas Precambrian rocks leave considerable to be desired for a conclusive tie with the Keweenawan belt. The volcanics are mostly rhyolite and not basic varieties as in the Keweenawan series, and orogeny including acidic intrusions and some metamorphism appears to be indicated. This is not characteristic of the Keweenawan belt.

J. Tuzo Wilson (1956) has suggested that the sediments of the Keweenawan, Huronian, and Mistassini groups along the Grenville front in Ontario and Quebec have been derived from the Grenville orogenic belt, and that a secondary mountain belt has resulted by their deformation at a later time. The Huronian rocks in Minnesota, Wisconsin, and Michigan have a much wider distribution than the Keweenawan series with its flanking faults, and are not so clearly a narrow belt as the Keweenawan. The writer sees in the Keweenawan belt one somewhat like the Triassic basins of the Piedmont crystalline province of the greater Appalachian mountain systems. See Chapter 9. These are long narrow fault-formed basins filled with thick sections of continental clastic sedi-

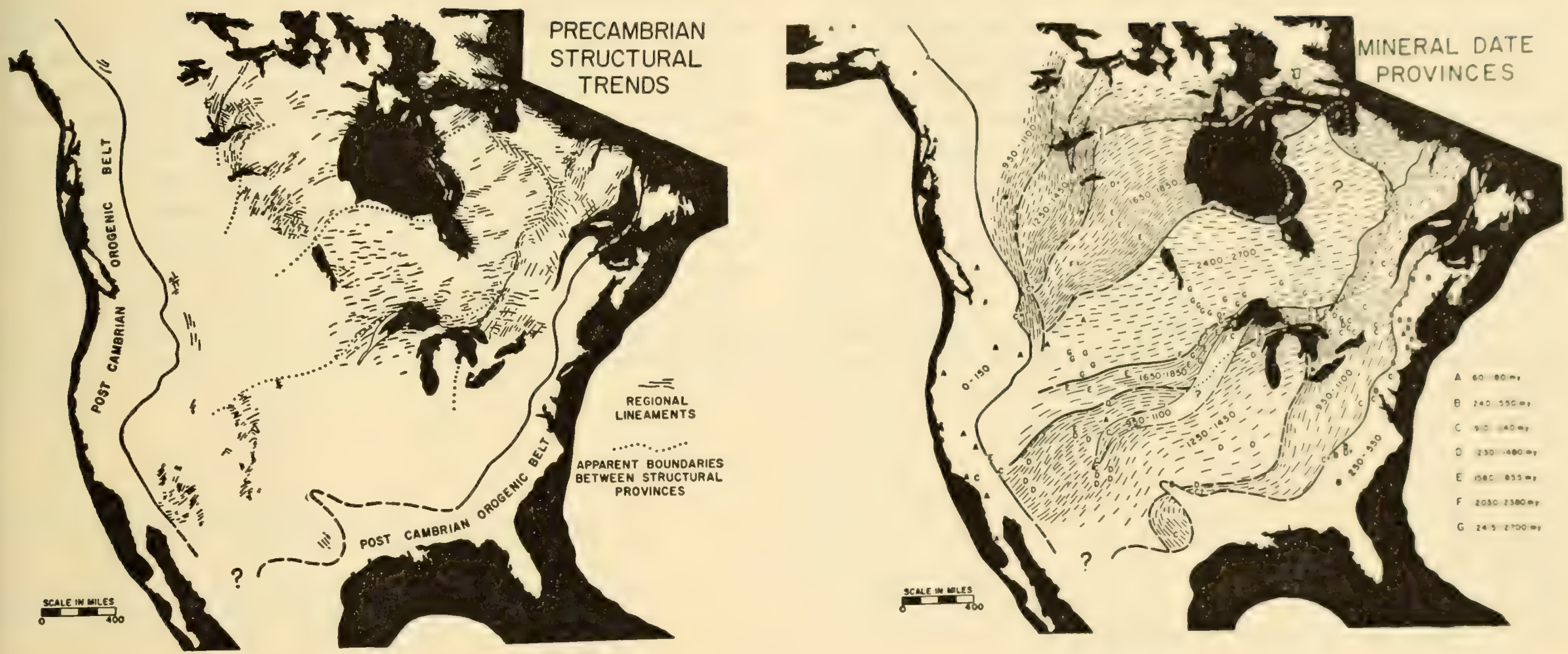


Fig. 4.9. Precambrian structural trends (left map) and mineral date provinces (right map) of North America. Reproduced from Gastil, 1960.

ments and basic flows, sills, and dikes. The basalts have been described as tholeiitic in both belts (Turner and Verhoogen, 1951). The significance of tholeiitic basalt is discussed in Chapter 33, and the occurrence is believed to be evidence that the belts formed under similar tectonic settings. Both are on the inside (toward the continent) of master orogenic belts involving extensive metamorphism and great batholithic intrusions. According to this interpretation the Keweenawan belt should mark approximately the inner front of the Grenville orogenic belt or province.

Regarding the succession in Texas, it is possible that the Swisher gabbroic terrane and parts of the Wichita igneous terrane are Keweenawan equivalents, and that the metasedimentary and volcanic (rhyolite) terranes are Huronian or somewhat older than the Keweenawan.

Texas Precambrian Rocks

In Texas and southeastern New Mexico a subsurface study of well samples penetrating the Precambrian has enabled Flawn (1956) to delimit several rock assemblages, which he calls terranes (Fig. 4.8). The basement rock is a granite dated about 100 m.y. old, and this is overlain

apparently unconformably by metasedimentary and volcanic rocks, and in one place by a gabbro sheet (?). The granitic intrusion therefore, correlated in age with the Grenville and Piedmont orogenies, and the metasediments and volcanics presumably with the Keweenawan series of the Lake Superior region. The Mazatzal orogenic belt appears to separate the Texas Precambrian assemblage from the Grenville, and hence, the most natural tectonic tie of the Texas assemblage appears to be with the Piedmont (Fig. 4.3).

Crystalline Piedmont

A broad belt of crystalline rocks extends from Alabama and Georgia northeastward along the Atlantic margin of the continent to New Jersey, and its relation to the Appalachian Mountains will be explained in some detail in Chapters 8 and 9. In summary, its rocks are now believed to be Precambrian and early Paleozoic in age, and to have been metamorphosed and intruded particularly during the Taconic and Acadian orogenies of Late Ordovician and Late Devonian ages, respectively. Age determinations on the rocks of the Piedmont indicate two ages, namely, an older one of Grenville age and a younger one of Paleozoic age. In fact, in one sample the zircon grains yielded an age of 1050 m.y., and the feldspars an age of 300 ± m.y. (Wetherill *et al.*, 1959). It is reasoned that this means an early orogeny in which the zircons were created, and a late orogeny in which the feldspars were formed but the zircons of the early orogeny left unaltered.

The distribution of dates so far published is shown on Fig. 4.3 and a comprehensive compilation and interpretation of Precambrian trends and orogenic belts of North America by Gastil (1960) is reproduced in Fig. 4.9.

5.

CENTRAL STABLE REGION OF THE UNITED STATES

GENERAL CHARACTERISTICS

The Central Stable Region of the United States is made up of a foundation of Precambrian crystalline rock previously described, with a veneer of sedimentary rock. The veneer varies greatly in thickness from place to place. For the most part, the Central Stable Region has suffered vertical movements, and broad basins and arches have formed. Some of the basins have more than 10,000 feet of strata in them, and in the cores of some of the arches the Precambrian crystalline rock is exposed. Some of the arches and sharper uplifts are not expressed in the surficial layers and have been revealed only by drilling operations. The arches, basins, and other structures of the Central Stable Region, with few exceptions, formed during the Paleozoic era, and many of them yield evidence of a prolonged history of development.

Up to Pennsylvanian time, there was a certain bilateral symmetry to the stable region, with a great medial transcontinental arch, and basins and smaller arches on either side. An approximate parallelism of a series of arches with the Ouachita and Appalachian orogenic belts was existent and is still apparent today.

During Mississippian, Pennsylvanian, and Permian time, great overlaps on some of the arches occurred. Others were either not completely buried or have since been partially exhumed by erosion. In some areas the Triassic overlapped on the Central Stable Region beyond the limits of the Permian, and especially in late Cretaceous time did epeiric seas extensively invade the region of arches and basins.

The large arches and basins are rippled and checked with numerous folds and faults; and these, with the unconformities created by the great overlaps, constitute immensely valuable structures for oil and gas accumulation. The strata also contain great coal deposits and numerous other nonmetallic mineral resources. Each basin and each arch will, therefore, be considered separately. The geologic and tectonic maps of Chapter 3 will be especially helpful in relating the diastrophic histories of the various major structures, and should be referred to repeatedly.

PRE-DEVONIAN BASINS

The basins of greatest extent and deepest subsidence in early Paleozoic time were the geosynclines along the western and eastern margins of the continent. Each constitutes an important part of our continent and will be discussed in separate chapters: the Paleozoic Cordilleran geosyncline in Chapter 6, and the Appalachian geosyncline in Chapters 7, 8, 11, 12, and 13. Refer to the map of Plate 2, Chapter 3, in the following paragraphs.

The Appalachian geosyncline subsided most in West Virginia, Virginia, Tennessee, and Alabama. In a small area across the border of Virginia and Tennessee, sediments accumulated to a thickness in excess of 25,000 feet during Cambrian, Ordovician, and Silurian time. A distinct sag in the

form of an embayment from Texas into southern Oklahoma resulted in the local accumulation of more than 6000 feet of strata, the chief formation of which was the Arbuckle limestone. Another embayment possibly extended to the western Texas region, where later the Pecos Range developed. The pre-Devonian sediments are thin in the Marathon and Ouachita systems as compared with the Appalachian system.

A rather deep basin formed in Michigan, Indiana, and Illinois in pre-Devonian time, approximately parallel with the Appalachian geosyncline. Its largest and deepest part is the present Michigan basin.

The great western geosyncline of early Paleozoic time extended from Alaska to southern California. It sank 15,000 to 20,000 feet across Nevada, and at the Nevada–California boundary it contained over 20,000 feet of beds (Nolan, 1943). No information is available farther southwest in California because of the extensive Mesozoic and Cenozoic cover, intensive metamorphism, and widespread Jurassic intrusions. The south termination of the geosyncline shown on the map is, therefore, hypothetical. The inner trough of the geosyncline becomes progressively deeper to the southwest and undeniably heads into the later Jurassic orogenic belt, which with still younger tectonic elements determines the margin of the continent today.

TRANSCONTINENTAL ARCH

General Features

During Devonian and Mississippian time the great Central Stable Region of North America consisted of three major divisions, a central northeast-southwest-trending Transcontinental Arch, and large basins, shelves, and arches and domes of various sizes on each side (Plates 3 and 5). The arch had three peninsular extensions to the southeast, one into Kansas and Missouri, the Ellis and Chautauqua arches and the Ozark dome; one into Wisconsin, the Wisconsin dome; and possibly one into Texas. It is also known to have sagged below sea level in two places where thin lower Paleozoic sediments were deposited, one in Colorado and one in Arizona. Until the rise of the ranges of the Ancestral Rockies and the Wichita systems, the Transcontinental Arch and its flanking basins dominated the landscape. The Transcontinental Arch may have bifurcated north of Lake Superior, with one arm extending northward on the west side of the Hudson Bay basin, and the other extending first eastward and then northward along the east side of the basin. This supposition is based on present Precambrian exposures, but paleontological evidence and newly found erosional outliers suggest that much of the area of the arms may have been submerged in early Paleozoic time.

Northeast of Colorado

The arch in Nebraska, South Dakota, and Minnesota was recognized by Schuchert and called Souxia. It was later clearly depicted by Levorsen (1931, Pl. 1), and then still later mapped by Ballard (1942). The boundaries of the formations shown on the geologic maps of the close of the Devonian and the close of the Mississippian are those preserved under the extensively overlapping Pennsylvanian strata (Plate 7) which covered most of the arch. Ballard has gathered together the available well records of the area and believes enough data is at hand to establish definitely the existence of the arch and fairly well the formational contacts on either side of it.

The arch was referred to as the continental backbone by Keith (1928) in his notable paper on "Structural symmetry of North America," and later, also, by Levorsen. The name implies that it was a strong, resistant, centrally located tectonic element with flanking basins and marginal orogenic belts in bilateral symmetry. With the exception of the peninsulas and sags previously mentioned, the bilateral symmetry of the United States part of the continent in a northeast direction was pronounced until the Pennsylvanian transgression. The building of the Ancestral Rockies altered conspicuously the aspect of the Transcontinental Arch, and then the late Mesozoic and early Cenozoic mountain building disturbances left the southwest half unrecognizable on a geologic map of the present time.

The Transcontinental Arch appears very dominant on a pre-Pennsylvanian geologic map, but this appearance should not be misinterpreted. During the Devonian and Mississippian, the arch was very low-lying and furnished chiefly chemical sediments to its flanking basins (Weller, 1931).

The "backbone" was also not very strong in resisting deformation. In its southwestern part, as previously mentioned, it was the site of Pennsylvanian and Cretaceous–Tertiary mountain building, and its other parts have been almost completely covered by Pennsylvanian, Permian, Mesozoic, and Cenozoic strata, in places of considerable thickness.

Wisconsin Dome

The area of central Wisconsin was probably uplifted several times in the Paleozoic, but evidence both for time and spatial relations is scarce and, therefore, all the geologic boundaries cannot be definitely fixed. The isopach maps of the *Ninth Annual Field Conference* of the Kansas Geological Society have been used as the chief source of information in making the interpretations shown on the maps of this book. The isopach maps generally show the existing thickness of the various formations or groups, and their compilers say that the original thickness and extent over the Wisconsin dome area is not certain. However, some of the formations thicken basinward under cover of protecting formations, and such contacts can be projected and the limits before burial located approximately.

Two pre-Devonian times of significant uplift are recognized; the first preceded the deposition of the St. Peter formation in Early Ordovician time, and the second followed the deposition of the Silurian beds. During the second uplift, an arch was formed that extended southeastward from Wisconsin into Illinois, almost to the city of Kankakee (Fig. 220, *Ninth Annual Field Conference*, Kansas Geological Society).

By the close of Mississippian time, a pronounced dome had appeared (Plate 6). A strip of Cambrian sediments extending southwest from the Keweenaw peninsula of Michigan indicates that the dome was separated from the Transcontinental Arch by a fairly broad, gentle syncline. A broad, noselike uplift extended southeastward from the Wisconsin dome in approximately the position of the post-Silurian arch and connected with the Kankakee arch of Illinois and Indiana (Plate 6). How far the Mississippian sediments spread over the dome area is not ascertainable, but following the late Mississippian uplift they were eroded back appreciably.

Colorado and Arizona

The rise of the Ancestral Rockies in late Mississippian and Pennsylvanian time destroyed the Transcontinental Arch in Colorado. The pre-Pennsylvanian sediments present are very thin, and cover the arch through central Colorado in a zone 100 miles wide. The zone was evidently the site of a gentle sag in the arch normal to its length, and as Burbank's (1933) map shows, it lines up almost precisely with the Wichita trough that others have shown in Oklahoma and Kansas. It seems, therefore, that the Wichita trough extended northwestward toward the Colorado sag, and not in the direction of the Amarillo Mountains in the Panhandle of Texas as has been suggested by some writers.

Arizona was mostly above water during the early Paleozoic (Stoyanow, 1942). The Mazatzal orogeny of Precambrian time (see previous discussion in this chapter) produced a chain of mountains that extended from southwestern Arizona to southwestern Colorado with subparallel folds and thrust faults trending northeastward (Huddle and Dobrovolny, 1950).

> The orogeny and associated intrusions took place after the Mazatzal quartzite was deposited. The mountains subsequently were well worn down by erosion, but the very resistant Mazatzal quartzite formed ridges along the core of the old mountain chain. The ridges served to separate the basins in which the rocks of the Apache and Unkar groups were deposited. . . . Both were considerably eroded before the Troy quartzite and Tapeats sandstone of Cambrian age were deposited. . . . After the deposition of the Cambrian sandstones, Mazatzal land probably was up-arched slightly and eroded, because the Martin formation in central Arizona rests on a surface of some relief. There are neither Ordovician nor Silurian rocks in central Arizona, and probably there never have been any. Cambrian rocks may have extended through the Mogollon sag, and a considerable thickness of them may have been removed from Mazatzal land during the long erosional interval between the retreat of the Late Cambrian seas and the spread of the Late Devonian seas. The gradual burial of the mountains and Mazatzal land before Pennsylvanian time is summarized diagrammatically in Fig. 4.4. Because the Martin formation was not deeply eroded prior to the deposition of the Redwall limestone, probably no diastrophic disturbances of Mazatzal land occurred at the close of the Devonian. After the Mississippian limestone was laid down, however, Mazatzal land again was uparched, as shown by the great erosional reduction

of the Redwall limestone on Mazatzal land and the related increase in the thickness of the red residual member of the Naco formation nearby (Huddle and Dobrovolny, 1950).

EASTERN INTERIOR BASINS AND ARCHES

General Features

Three basins of subsidence and sedimentation had become clearly established by late Devonian time southeast of the Transcontinental Arch, namely, the Michigan basin, the Illinois–Indiana–Kentucky basin (Eastern Interior basin), and the West Virginia–Pennsylvanian basin (Appalachian basin). In Pennsylvanian time a fourth became defined, which is called the Western Interior basin as a coal province, and the Forest City basin as an oil province. See map, Fig. 5.1, and Plate 7. The Western Interior and Eastern Interior basins were first so labeled when studied as coal basins of Pennsylvanian age, and although the nomenclature is not consistent with the state name applied to the Michigan basin, also a coal basin, it is generally retained and used today.

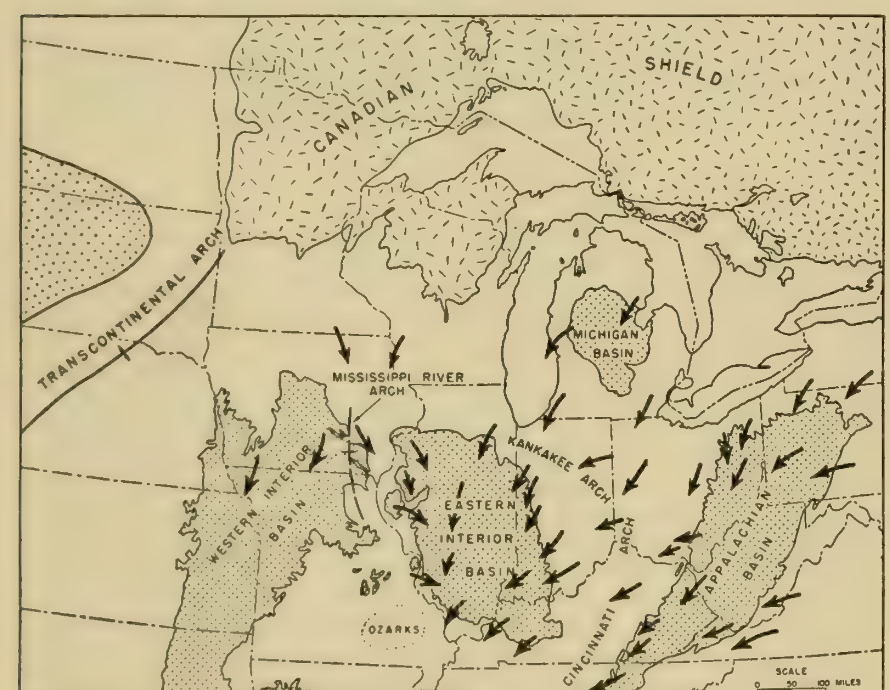


Fig. 5.1. Basins southeast of the Transcontinental Arch showing areas of sand accumulation in early Pennsylvanian time and the direction of stream transport. After Potter and Siever, 1956.

Appalachian Basin

The history of the Appalachian basin is recounted in Chapter 7 in connection with the Appalachian Mountains. As shown on the map, Fig. 5.1, it lies between the Valley and Ridge Province of the Appalachians and the Cincinnati arch, but in its development its deepest part lay in the mountainous belt; the eastern half of the basin became involved in folding and thrusting in late Paleozoic time leaving the western half relatively undeformed and what is now called the Appalachian basin. See Figs. 8.11 and 8.12. It is filled with a remarkable succession of miogeosynclinal and shelf strata ranging in age from Cambrian to Permian.

Michigan Basin

In pre-Devonian time, the Michigan and Illinois–Indiana–Kentucky basins were continuous; but beginning in the Devonian, the Kankakee arch began to form, and the two basins became increasingly individualistic thereafter. The Michigan basin today is circumscribed by the Great Lakes depressions on the west, north, and east, and by the Cincinnati dome on the south. It consists of a sequence of beds representative of all periods of the Paleozoic, cast in saucer fashion, each one of which is smaller than the preceding on which it rests. The youngest strata are thin and patchy red beds of either Upper Pennsylvanian or Permian age. All Paleozoic strata are overlain and nearly completely blanketed by a layer of glacial drift which ranges in thickness from a few feet to 1200 feet. As the basin subsided through the Paleozoic, its crystalline precambrian floor acquired the configuration shown in Figs. 5.2 and 5.3. The total thickness of sediments in the basin is about 14,000 feet (Cohee, 1948).

The major unconformity in the Paleozoic sequence is at the base of the St. Peter sandstone and the Trenton and Black River limestones. See Figs. 5.4 and 5.5. The St. Peter sandstone is late Lower Ordovician, and marks the time of uplift and erosion. When traced eastward from Indiana to Ohio and northeastward into Ontario in well logs, the Lower and Middle Ordovician formations rest successively across the several formations of the Upper Cambrian, and finally come to rest directly on the Precambrian crystallines of the Canadian Shield. Through western Ontario, the Cambrian beds are absent.

Significant units in the Michigan basin are the evaporite series of the Silurian and Devonian. A number of beds of salt are present throughout much of the basin and southwestern Ontario which in places may aggregate over 2000 feet in thickness. Porous dolomites in these evaporite series are reservoir rocks for oil and gas, and many oil fields have been developed in the basin. Very gentle folds or "highs" ripple the basin beds and take an irregular northwest-southeast direction. They have served to trap the oil (Fig. 5.4).

In the Straits of Mackinac region, the most prominent outcrops are a limestone breccia. It is noted for its resistance to erosion and forms the scenic pillars and cliffs of the region. The map of Fig. 5.6 shows its known distribution.

The columns of breccia, according to Landes (1945), may range up to 1500 feet in vertical dimension. The solution of Silurian salt has resulted in subsidence and roof collapse, and the breccias are the result. Certain blocks can be shown to have fallen or settled 600 feet. The formations involved and the nature of the breccias are illustrated in the cross section of Fig. 5.7. Supporting the salt solution and collapse theory is the map showing the abrupt thinning of the Salina salt in the Mackinac Straits region (Fig. 5.8). The solution of salt and the collapse of the overlying layers of limestone and dolomite took place chiefly in pre-Dundee time (Middle Devonian), but even now some leaching may be occurring.

Great Lakes Depressions

The Salina salt emerges from the basin in a horseshoe-shaped pattern that corresponds closely with Lake Michigan and Lake Huron. The out-

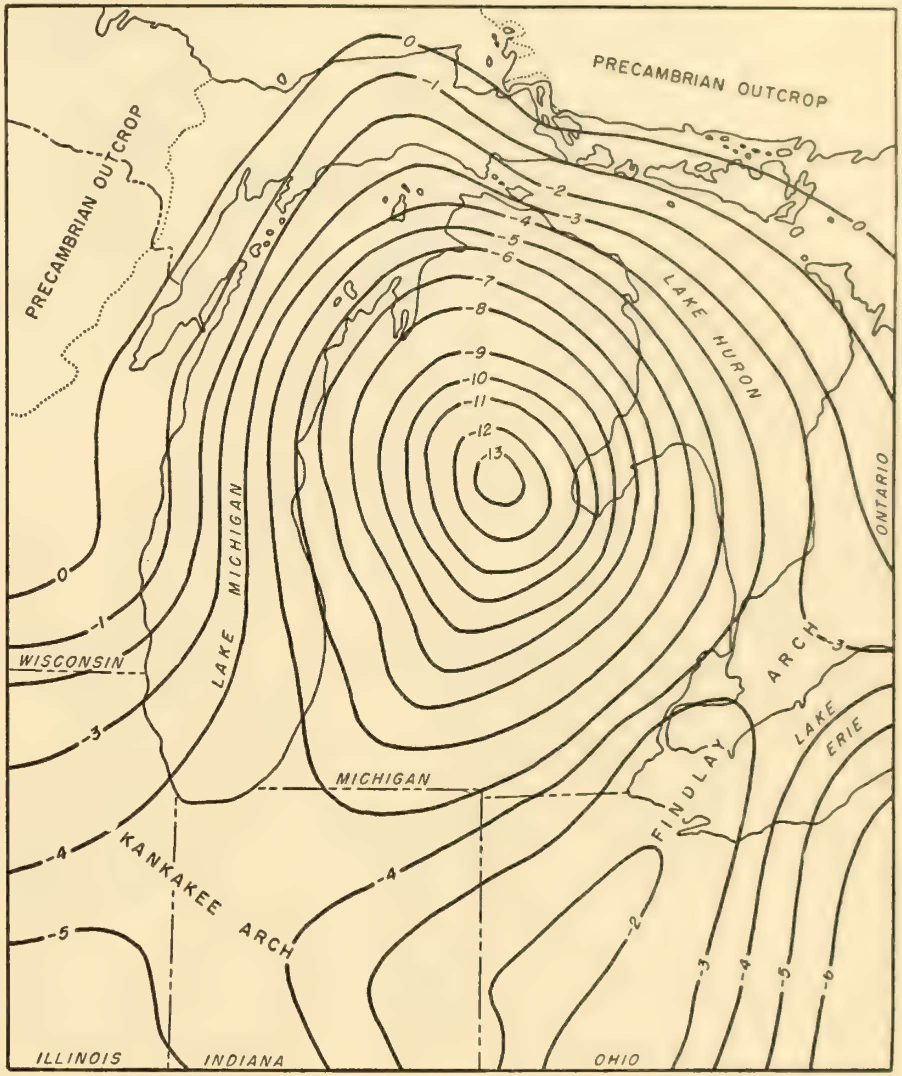


Fig. 5.2. Configuration of the Precambrian floor in the Michigan basin and adjoining areas. Contours in thousands of feet. After Cohee, 1948.

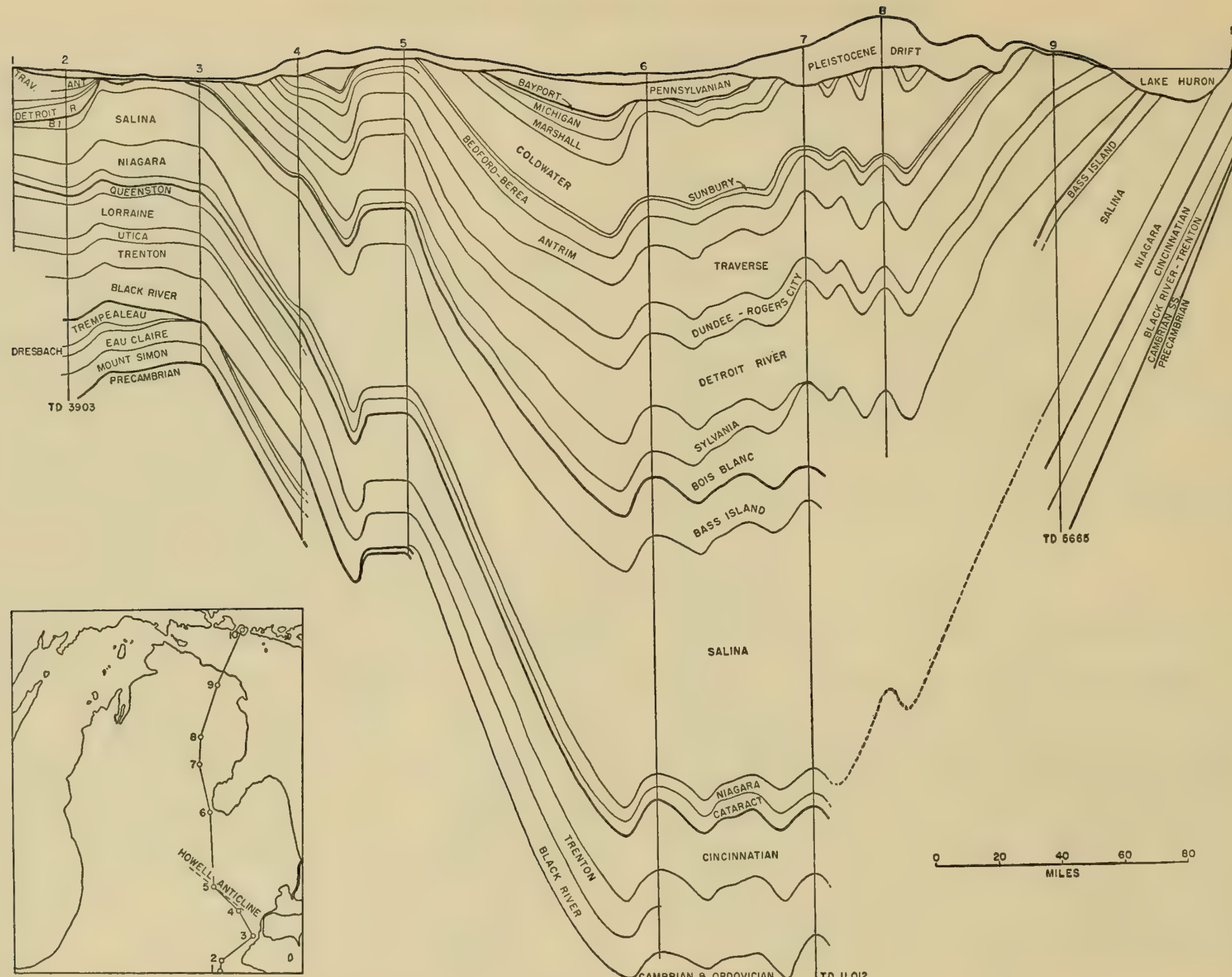


Fig. 5.3. Cross section of the Michigan basin, after American Association of Petroleum Geologists, 1954, *Geologic Cross Section of Paleozoic Rocks,* central Mississippi to northern Michigan. The Cambrian is mostly a sandstone and shale sequence; the Black River through Traverse a limestone, dolomite, and evaporite sequence, the Antrim through Michigan a shale and sandstone sequence.

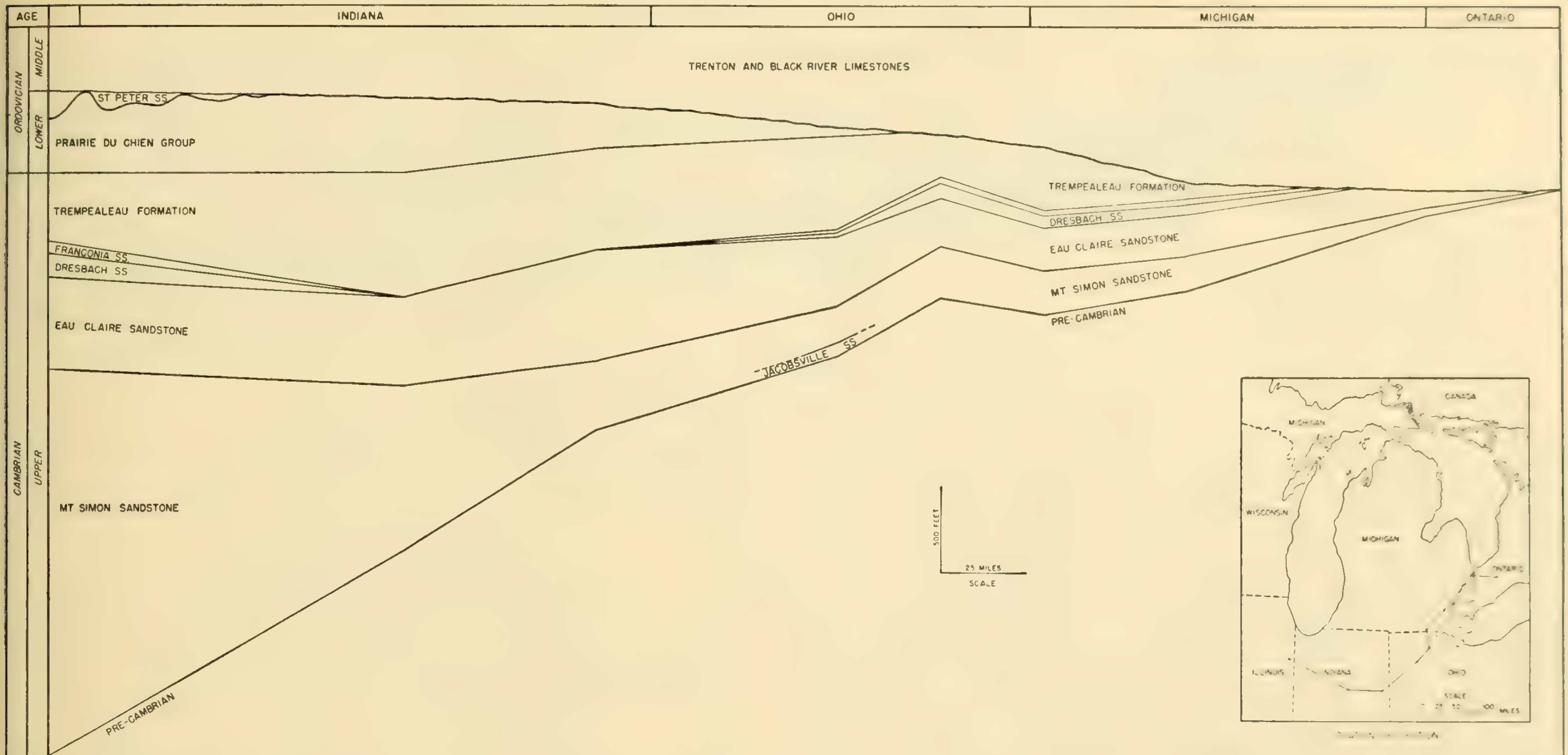


Fig. 5.4. Cross section from Illinois to western Ontario showing the unconformity at the base of the St. Peter sandstone and the Trenton and Black River limestones. Top of Trenton is taken as horizontal datum. Younger formations and present structures not shown. By George Cohee, U.S. Geological Survey.

crop then swings eastward through the basins of Lake Erie and Lake Ontario. The salt would emerge mostly under water, and since the aggregate thickness of salt beds that once may have cropped out was several hundred feet, it has been suggested (Newcombe, 1933) that the depressions of the Great Lakes (excepting Superior) may be due to salt solution and consequent subsidence. The basins do not correspond to faults or folds, and were probably existent long before the Pleistocene ice lobes occupied them. The theory of salt solution seems the most logical explanation yet advanced.

The Lake Superior depression is north of the belt of salt outcrop and is mostly in Precambrian rocks. The northwest shore may correspond to a fault, and the lake bottom topography suggests fault scarps. Because the

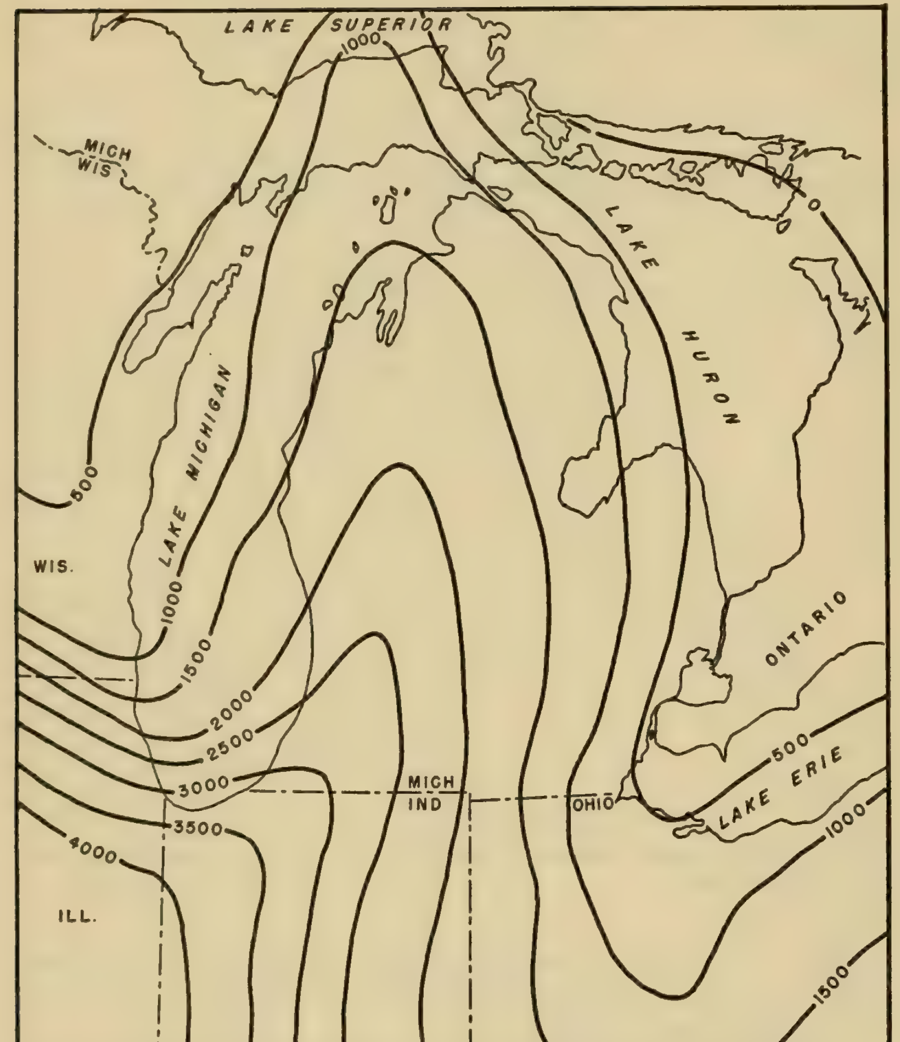


Fig. 5.5. Thickness of Upper Cambrian and Lower Ordovician Rocks in the Michigan Basin. After Cohee, 1948.

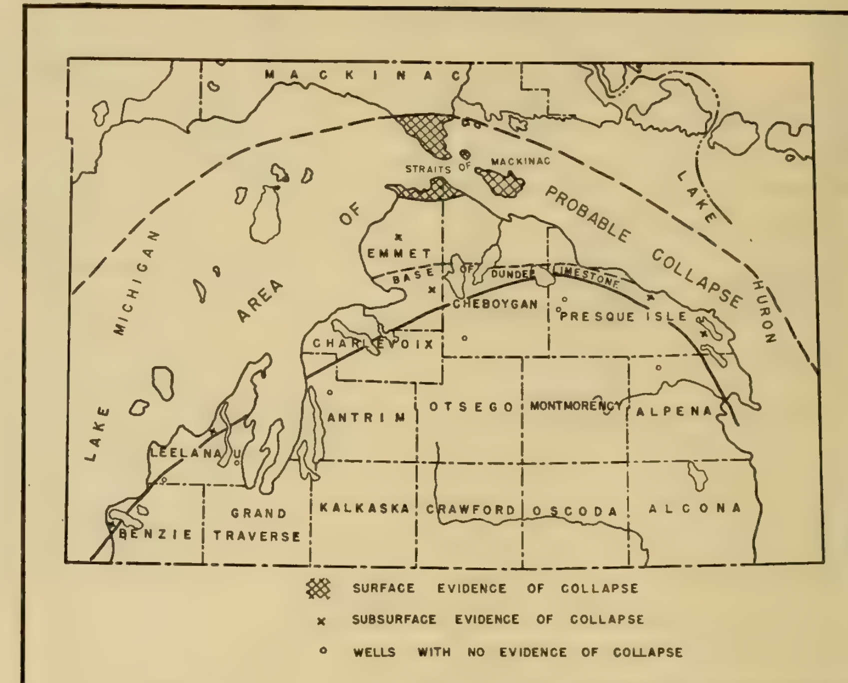


Fig. 5.6. Map of Mackinac Straits areas showing zone of collapse and exposures of Mackinac breccia. Reproduced from Landes, 1945.

faults have been regarded as either Precambrian or late Paleozoic in age, they are very ancient, and any scarps would be erosional features of the fault-line variety. Previous conjecture places the Grenville front in the position of the lake, and later subsidence along this zone may have occurred to form the lake basin. It must be conceded, however, that the origin of the Lake Superior basin, over 1000 feet deep in places, has not yet been worked out satisfactorily.

Eastern Interior Basin

The Eastern Interior or Illinois–Indiana–Kentucky basin is deepest in Wayne, White, and Hamilton counties where the base of the Mississippian

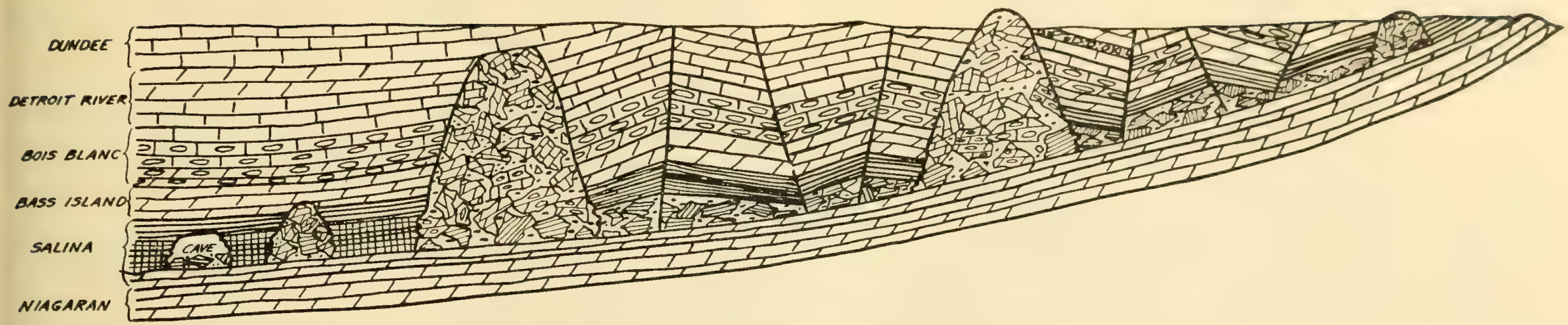


Fig. 5.7. Hypothetical section of the Mackinac Straits region showing collapse formations above the Niagara limestone, and the breccia chimneys and stacks. Reproduced from Landes, 1945.

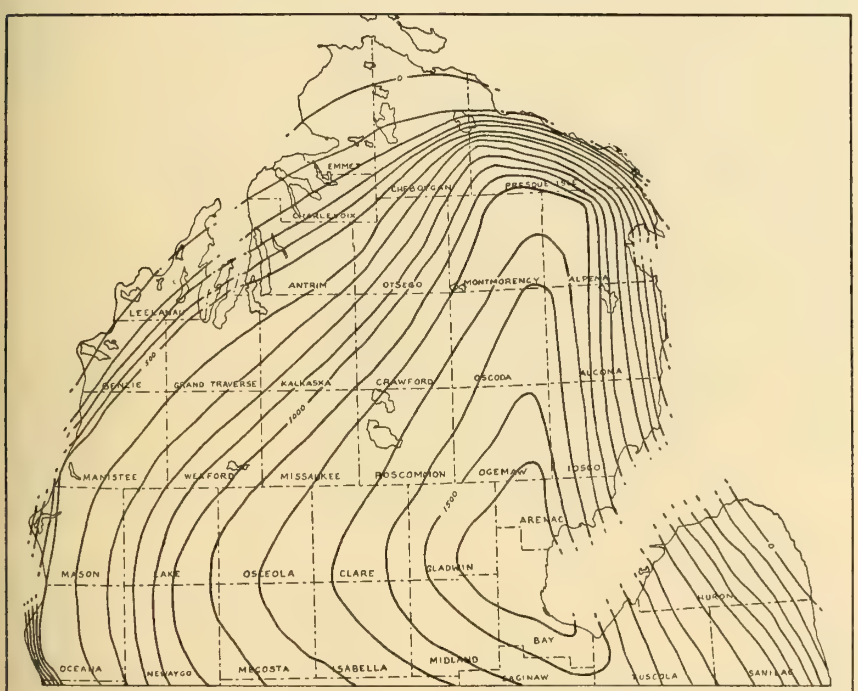


Fig. 5.8. Isopach map showing aggregate thickness of Salina salt. Reproduced from Landes, 1945.

shales is 4800 feet below sea level. As previously explained, the Eastern Interior basin was part of a depression that included the Michigan in pre-Devonian time, but from then on the two basins sank separately, leaving the Kankakee arch between.

The La Salle anticlinal belt (see Fig. 5.9) is a row of anticlines arranged *en echelon,* and it extends over 200 miles from north central to southeastern Illinois. The north end of the *en echelon* belt may be connected with the east-west trending Savanna–Sabula anticline (Eckblaw, personal communication), which extends into eastern Iowa. The south end may merge with the Wabash River anticline. The La Salle anticlinal belt formed chiefly during the Pennsylvanian period and divided the pre-Pennsylvanian Illinios–Indiana–Kentucky basin into two parts, the larger and western of which is generally known as the Illinois basin. The Oakland anticline borders the La Salle closely on the east.

The first deformation took place in post-Chester, pre-Pennsylvanian time (Fig. 5.10). Further deformation continued during the Pennsylvanian progressively southward. In La Salle and Douglass counties at the north end, the early movements were the greatest, and the crest of the anticline was elevated 900 to 1400 feet above the adjacent basins. In Lawrence and Wabash counties to the south, the greatest movements occurred within the Pennsylvanian. Since the Pennsylvanian beds are

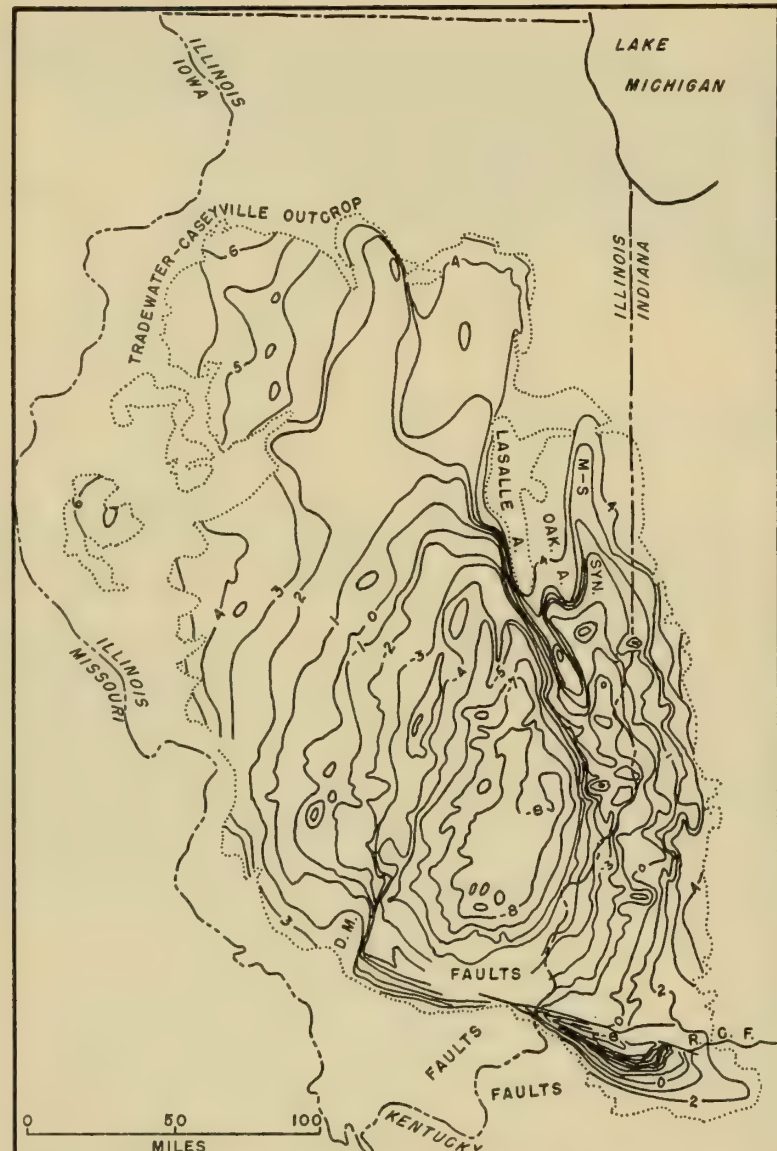


Fig. 5.9 Structure contour map of Eastern Interior basin. Contours on Illinois Coal No. 2, in hundreds of feet. After Wanless, 1955. Oak. A., Oakland anticline; M-S Syn., Marshall-Sidell syncline; D.M., Duquoin monocline; R.C.F., Roush Creek fault zone.

slightly folded over the anticline, it is possible that some movement occurred after they were deposited as well as during the time of deposition.

At the beginning of the Pennsylvanian there was a regional southwest slope furrowed by numerous subparallel valleys as deep as 200 feet. An eastward slope prevailed along the western border of the basin with

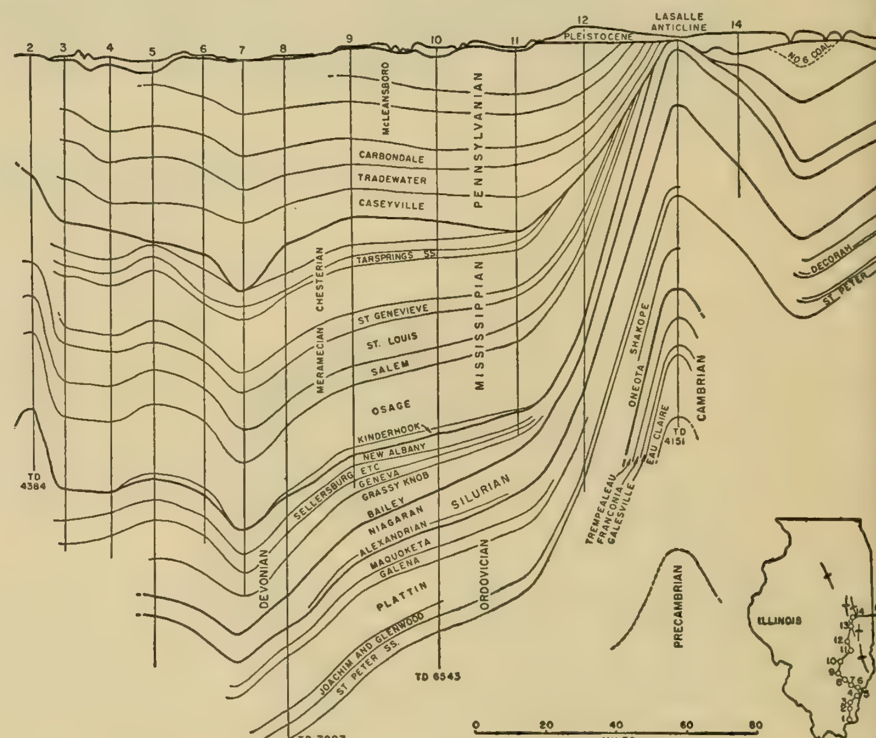


Fig. 5.10. Cross section of the Illinois basin, after American Association of Petroleum Geologists 1954, *Geologic Cross Section of Paleozoic Rocks,* central Mississippi to northern Michigan. The Eau Claire and older beds of the Cambrian are sandstone and shaly sandstone; from the upper part of the Eau Claire through the Ordovician, Silurian, Devonian, and the Mississippian to the Upper Mississippian Chester series the sequence is dominantly limestone and dolomite with much chert. The St. Peter is conspicuous sandstone in the Ordovician, and the Osage of the Mississippian has considerable sandstone and shale toward the La Salle anticlinal belt. The Chester and Pennsylvanian strata are sandstone and shale with several thin limestone beds, and coal in the Pennsylvania.

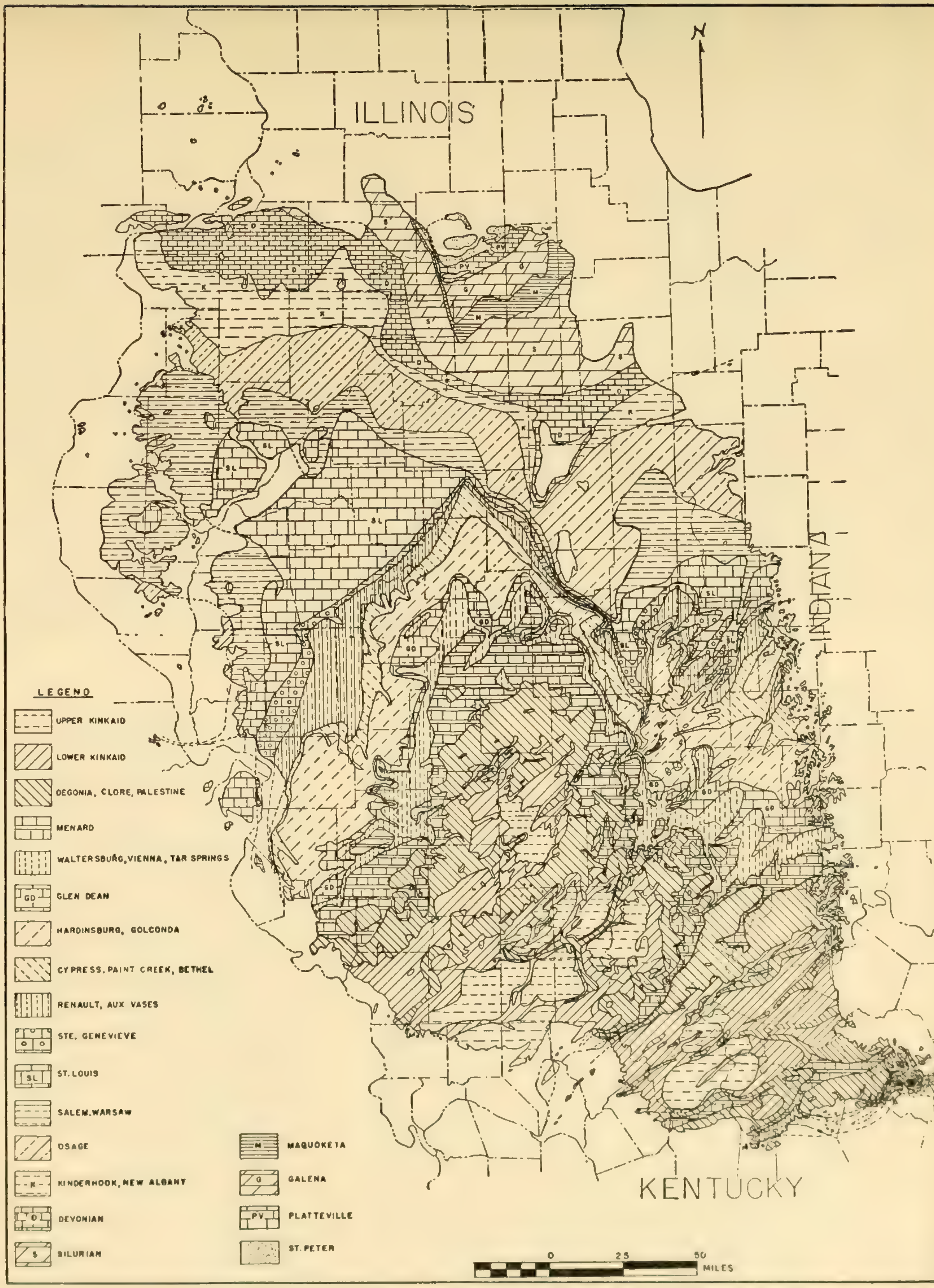


Fig. 5.11. Pre-Pennsylvanian paleogeologic map of Eastern Interior basin. Reproduced from Wanless, 1955.

smaller valleys. The geologic map at this time would have appeared as in Fig. 5.11, when formations from the Middle Ordovician St. Peter sandstone to the Upper Mississippian Kinkaid limestone cropped out. The southeast border of the basin sank progressively and resulted in a regular increase in thickness of the uppermost Pennsylvanian strata in that direction (Wanless, 1955).

The assemblage of faults in southern Illinois, shown on the map of Fig. 5.9, consists principally of the following trends: (1) the east-west trend of the Rough Creek–Shawneetown system which extends west into Illinois as the Cottage Grove and associated faults; (2) a prominent northeast-southwest system of faults which is dominant in the fluorspar district; and (3) the Wabash Valley fault system with north-northeast trend, a few of which cross and offset the Rough Creek fault. These faults are post-Pennsylvanian in age and the maximum throw is about 800 feet along the Rough Creek fault.

Studies of crossbedding and stratigraphic relations indicate that the late Mississippian Chester sands as well as those of the Pennsylvanian came mostly from the northeast (Potter *et al.*, 1958, Potter and Siever, 1956, and Wanless, 1955), and some were probably carried by streams from the site of the Michigan basin across the site of the previous Kankakee arch. A minor amount of sand came from the Transcontinental Arch. See Fig. 5.1.

Nashville Dome

The Nashville dome is at present the site of a topographic basin, with surrounding escarpments of successively younger rocks. Ordovician strata are the oldest rocks exposed in the core, and the escarpments are in the overlapping Mississippian and Pennsylvanian formations. The dome experienced several movements in pre-Chattanooga (early Mississippian) time, synchronous with those of the hinterland of the Appalachian geosyncline, according to Wilson (1935). The dome was below sea level during several epochs of various lengths of time, and during other times the central part was above sea level but probably so slightly emergent that little erosion occurred. The structure is a broad, gentle arch, less because of uplift than because of greater subsidence of the adjacent basins. Its domal structure was acquired by gentle sags between it and the Ozark dome on the west (Wilson, 1939) and the Cincinnati dome on the north (MacFarlan, 1943). See cross section of Fig. 5.12.

The first major uplift in which considerable truncation of the beds occurred was in late Devonian time. The Chattanooga shale rests on the Trenton (Ordovician), showing that about 500 feet of beds had been eroded away in the central part of the dome consequent to this pre-Mississippian doming (Wilson and Born, 1943).

The second major uplift was in late Mississippian and early Pennsylvanian time, when its associated domes, the Ozark and Cincinnati, were also elevated (Plate 5). The Chattanooga shale was domed gently, producing regional dips of 16 feet per mile on the flanks, and along the axis, both northeast and southwest, of about 8 feet per mile. A structural relief of 700 feet was acquired by the dome above the saddle separating it from the Cincinnati dome on the north. The structural relief of the dome over the flanking basins was at least twice as much (Wilson and Spain, 1936).

Detailed structure contour maps reveal many local irregularities in the Nashville dome. A conspicuous "grain" to the northwest is noted by Wilson and Born (1943), and axes of folds may be drawn in a few places. A structure contour map of the Pencil Cave (Ordovician) formation shows the grain equally as well as one drawn on the Chattanooga shale (Mississippian), but the local structures are not closely superposed. It may, therefore, be inferred that part of them originated in pre-Chattanooga time, and part in post-Chattanooga.

Cincinnati Dome

The Cincinnati dome is much like the Nashville dome, and is separated from it by a shallow structural saddle. Several writers refer to the two domes together as the Cincinnati arch, with the central part of the northern structure, the Jessamine dome, and the Nashville dome as elements of it. The Cincinnati dome splits into two branches on the north, one extending to the west-northwest and the other to the north-northeast, which are known, respectively, as the Kankakee arch and the Findlay arch.

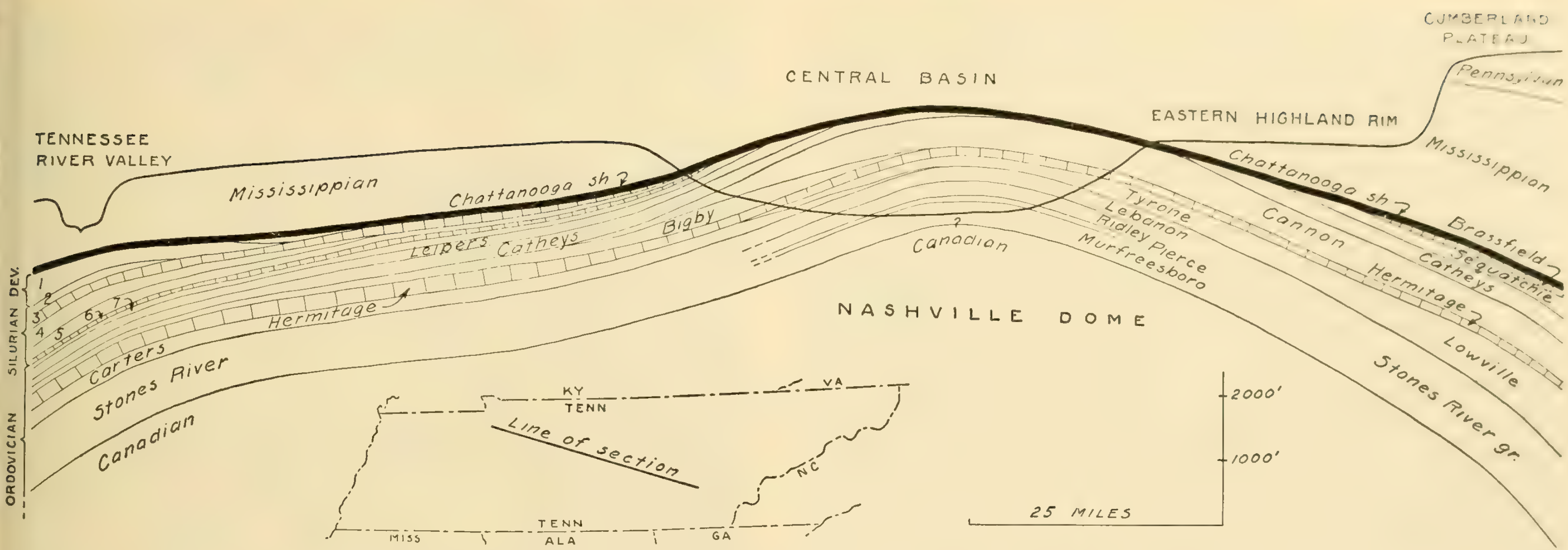


Fig. 5.12. Section across the Nashville dome, after C. W. Wilson, Jr., 1935. 1, Lower and Middle Devonian; 2, Decatur; 3, Lobelville; 4, Beach River, Bob, and Dizon; 5, Lego, Waldron, Laurel, and Osgood; 6, Brassfield; 7, Richmond.

The Cincinnati dome probably had an early Paleozoic history much like the Nashville dome, but the first elevation in which appreciable erosion occurred preceded slightly the one in the Nashville dome. MacFarlan (1943) shows that the Middle Devonian (Boyle) limestone overlaps successively older formations toward the center of the dome where it rests on the Ordovician (Richmond and Maysville). The Lower Mississippian shale (Ohio shale, probably the Chattanooga equivalent) has been found to "cut out" the Boyle limestone in a few places, and therefore locally some late Devonian movement has been suggested.

Preceding the Mid-Devonian uplift of the Cincinnati dome and about 100 miles east of it, arose the Waverly arch in Early Ordovician time. It has a structural relief of 750 feet (Woodward, 1961).

The Pennsylvanian–Mississippian contact is one of marked disconformity and one of considerable relief as shown in a number of Pottsville-filled valleys. The post-Mississippian uplift represented by the unconformity was much broader than the doming of Middle Devonian time. Compare Plates 5 and 6. It is generally regarded that after the late Mississippian arching, the Cincinnati dome was submerged, and that Pennsylvanian beds from the Appalachian region spread westward across it so that the Appalachian and central interior coal fields were connected. Several of the conglomerates, fireclays, and limestones have been correlated across the dome. See Plate 7.

In order to produce the present distribution of the Pennsylvanian strata, still another broad, gentle arching is required in post-Pennsylvanian time. This is shown on the tectonic map of Plate 8.

Some faults cut the dome, and these will be described later as part of

a large fault zone that extends across several states. Local structures are not as well mapped as in the Nashville dome; but as far as known, the perceptible northwest "grain" does not exist. Instead, one or two "highs" have been described on the eastern flank of the dome that trend parallel with the main axis. It may be that with better contouring, a northwest direction of local structures will be noted.

Kankakee Arch

The Kankakee arch, as defined by Ekblaw (1938), is the northwest branch of the Cincinnati dome, and passes in a northwest direction across Indiana and Illinois, connecting with the Wisconsin dome. Kankakee is preferred to Wabash, a name sometimes used. The earliest significant uplift preceded the deposition of the St. Peter sandstone, as in the Wisconsin dome. The St. Peter sandstone rests on Cambrian beds at Oregon, Illinois, indicating arching above sea level and removal of 500 to 600 feet of rock in this early movement. The Cambrian and Prairie du Chien (pre-St. Peter) beds are believed to be about 4000 feet thick, both on the Kankakee arch and in the Illinois basin, and therefore the arch was evidently an area of subsidence just as much as the basin until Early Ordovician time.

Oil wells show that the structural relief at present, if measured on the top of the Trenton limestone, is about 6000 or more feet in relation to the Illinois basin and 10,000 feet in relation to the Michigan basin. As the Trenton is above the St. Peter, the arch has acquired this much additional structural relief since the pre-St. Peter uplift. It is clear that the large part of this structural relief is a result of subsidence of the basins on either side of the arch, and that the upward movements of the arch itself, sufficient to cause it to be eroded, contributed only in small part to the relief. See Figs. 5.4 and 5.5 for pre-St. Peter structural relations.

The only reflection of the Middle and Late Devonian uplifts of the nearby Cincinnati and Nashville domes is the conspicuous thinning of one of the zones of the Traverse group in the Michigan basin toward the arch (Cohee, personal communication). The greater subsidence of the basin area than the arch area, as indicated by this zone in the Traverse, occurred in late mid-Devonian. The basin had previously sunk rapidly, and a thick evaporite series was deposited during the Silurian and pre-Traverse Devonian. These thick salt, gypsum, limestone, and dolomite beds are represented by thinner nonevaporite series in the Illinois basin, and hence the structural relief of the arch is not so great to the southwest as to the northeast.

The early Mississippian seas probably spread over the arch even though Lower Mississippian rocks are not there today. This is concluded because the beds do not display any characteristics of overlap on a land area. The Upper Mississippian (Chester) beds of Illinois are not represented in the Michigan basin, nor anywhere north of the Kankakee arch, and it therefore seems that in late Mississippian time the arch and the area to the northeast were gently emergent, and from this region and still farther north the Chester sands were derived. Since the present structure displays the geologic pattern of a broad anticline with Silurian rocks in the core and Devonian and Mississippian successively away on either side, it follows that in addition to regional uplift over the Great Lakes region in late Mississippian time there must also have been local uplift along the arch. This movement occurred at the same time as the one described in the Cincinnati dome with which the Kankakee arch merges.

The deposition of Pennsylvanian sediments across the Cincinnati dome on a surface of appreciable relief corresponds to the well-known Pennsylvanian overlap in Illinois south of the Kankakee arch and over the La Salle anticlinal belt. The Upper Mississippian and pre-Pennsylvanian uplift along the arch was probably a movement of only a few hundred feet. Again it was the appreciable subsidence of the adjacent basins that contributed most to the arch structure.

Because the Pennsylvanian strata were gently arched and eroded back from the Cincinnati arch, a post-Pennsylvanian uplift of gentle but broad dimensions is indicated. It appears that the uplift spread northward so as to embrace the Kankakee arch, the Wisconsin dome, the Michigan basin, and the southern part of the Canadian shield.

In summary, the Kankakee arch acquired its structural relief chiefly by greater subsidence of the basins on its sides than by actual uplift. It was lifted out of water in early Ordovician time, and in one place it suffered 600 feet of erosion. Again it rose out of water in late Mississippian time, and finally participated in a regional uplift of the Great Lakes region in the late Pennsylvanian.

A sag between Peru and Logansport across the arch is called the Logansport sag, and many other minor irregularities make up the oil field structures in the area.

Findlay Arch

The Findlay arch is the right arm of the Cincinnati dome, and extends north-northeastward into the peninsular area of Ontario and thence to the Canadian Shield (Plate 5). It is similar in size and relief to the Kankakee arch and, since the early Ordovician uplift, it has had a similar history (Cohee, personal communication). It was not an area where thick pre-St. Peter sediments accumulated, and may actually have been a low ridge of Precambrian rock at the beginning of Cambrian deposition (Cohee, personal communication).

The uplift along the Findlay arch was localized and of somewhat greater magnitude than along the Kankakee arch (Cohee, personal communication). The cross section, Fig. 5.4, shows the base of the Black River and the progressive overlap northward to the Precambrian crystallines of southeastern Ontario.

The names Lima axis and Sandusky arch (Phinney, 1891), Algonquin axis (Kay, 1942), and Cataract axis have been used for all or part of the arch, but Findlay arch is preferred by Ekblaw (1938) and others. A sag in the axis near Chatham, as contoured by Cohee (personal communication), reflects movements at the same time approximately as those in the arch. The cross structure is called the Chatham sag.

Arches of Central Kansas

The geologic map of mid-Pennsylvanian time (Plate 7) shows the superposition of one arch over another in central Kansas, with axes trending in slightly different directions. At the close of the Devonian, a broad arch, for which the name Ellis is reserved (Moore and Jewett, 1942), rose (Plate 5) and was eroded so that the Lower Ordovician Arbuckle limestone was exposed in the core. The Mississippian seas then lapped onto the Ellis arch and perhaps covered it. Post-Mississippian arching in a somewhat more northerly direction and in a narrower zone resulted in the erosion of the Mississippian strata and the exposing of the strata in the Ellis arch again. This new uplift is called the central Kansas arch. However, the local folds that developed parallel with the major axis of the Ellis arch trend obliquely across the core of the central Kansas arch. Examine the cross section of Fig. 5.13 and the map of Fig. 14.1.

The Ellis arch continued eastward as the Chautauqua to the Ozark dome. The Chautauqua connection existed only at the close of the Devonian.

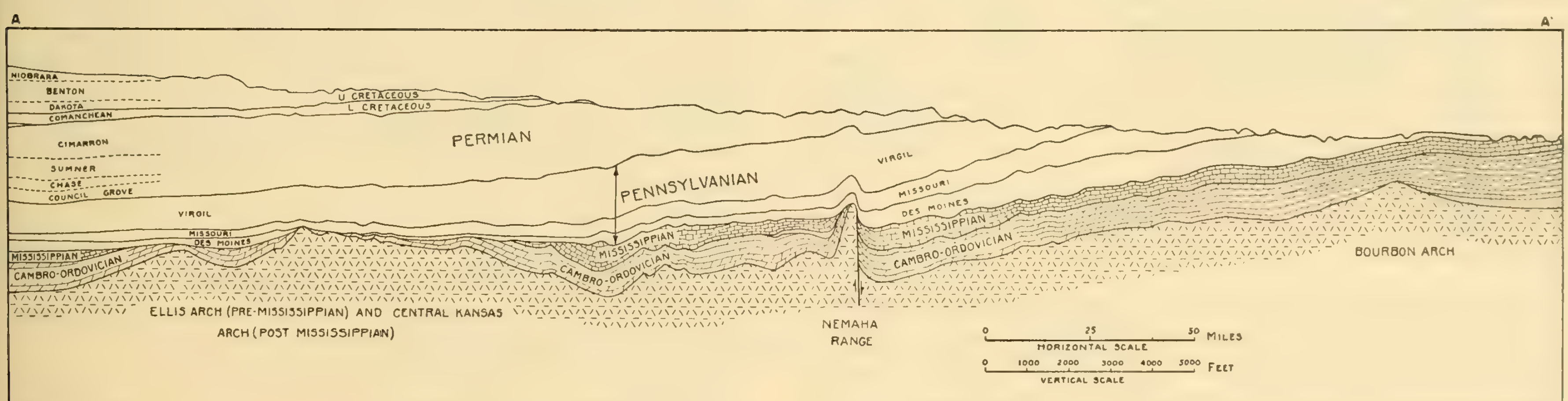


Fig. 5.13. Section along central Kansas arch, Nemaha Range, and Bourbon arch. taken from cross section by Betty Kellett (1932). Line of cross section shown on map of Fig. 14.1.

Nemaha Range

A very sharp uplift, the Nemaha Range, trends south-southwest from Omaha through southeastern Nebraska across Kansas into northern Oklahoma, but is now buried. See Figs. 5.13 and 14.1 and Plate 6. It came into mountainous relief during early Pennsylvanian time, because the Mississippian strata are tilted up and truncated along its sides. Uplift and dissection were sufficient to expose Precambrian crystalline rocks in the core before burial. See cross section of Fig. 5.14. Structural relief is 3600 feet in the central part of the range, and the east flank is so steep and straight that a block-fault movement has been visualized (Lee *et al.*, 1946). The range was eroded rapidly so that the Pennsylvanian strata, partly derived from the range itself, encroached on its flanks and, together with much exotic material perhaps in part from the early Ouachitas, finally buried the range. The present depth of the "granite" at the Kansas and Nebraska line is only about 400 feet (500 feet above sea level), but at the Kansas and Oklahoma line, it is over 3000 feet below the surface (2500 feet below sea level).

The Nemaha Range contrasts strongly with the central Kansas arch in relief and symmetry. The Nemaha Range has 3600 feet of relief, whereas the arch has 1500. The range has a very steep eastern front and gentle back slope, whereas the arch is symmetrical and gentle. The nearly north-south trend of the Nemaha Range is unlike the northwest trend of the broad, gentle arches, and this sets it apart from the arches as a different structural type. It resembles the Colorado Range of the Ancestral Rockies, and therefore the characterization of it as a range is more appropriate than as an arch, anticline, or ridge, as it has variously been called.

Bourbon Arch

Slightly north of the site of the previous Chautauqua arch, a later but narrower one rose in early Pennsylvania time. It was probably a shallow platform between the Forest City basin on the north and the Cherokee basin on the south (Moore and Jewett, 1942). See Fig. 5.13.

Ozark Dome

At present, the Ozark dome is a broad, nearly circular area of Cambrian and Ordovician limestones, surrounded by escarpments of Mississippian limestone. In the east central part, knobs of pre-Cambrian crystalline rocks project through the Cambrian and Ordovician strata to the surface. The crystalline outcrops occur in southeastern Missouri, the area of the St. Francis Mountains, and the strata dip everywhere away from them (Croneis, 1930). The dome itself spreads over two-thirds of the state, and also into northern Arkansas where the Boston Mountains make up the southern flank. The Precambrian surface had considerable relief, and the younger strata were deposited on it with initial dips in places up to 30 degrees (Bridge, 1930).

The first major unconformity in the Paleozoic succession around the Ozark dome, especially on the west side in the Forest City basin, is at the base of the St. Peter sandstone. Lee *et al.* (1946) summarize the subsurface geology in maps and cross sections and show that subsidence took place in the Ozark region in pre-St. Peter time, while upwarping took place in southeastern Nebraska and northeastern Kansas (the Nebraska arch). The structural relief between basin and uplift was about 2000 feet.

With the coming of St. Peter time, the crustal movements were reversed, and the Ozark basin now started to rise as the Nebraska arch started to subside. At the end of Silurian time, widespread erosion occurred, with the greatest amount around the rising Ozark dome. The Devonian strata not only rest on the truncated Silurian and older rocks around the dome, but themselves in turn are truncated and covered by the Mississippian strata.

The Mississippian overlap is most extensive and very well known from many well records on the west side of the dome. Consult the geologic map of the close of the Devonian, Plate 5, and cross sections of Figs. 5.15 and 5.16. The unconformity indicates that the dome was again uplifted in late Devonian time and considerably eroded. The pre-Mississippian geologic maps of the region (Moore and Jewett, 1942; Wrather, 1933; Lee *et al.*, 1946), together with surface outcrops, indicate that the Ellis arch,

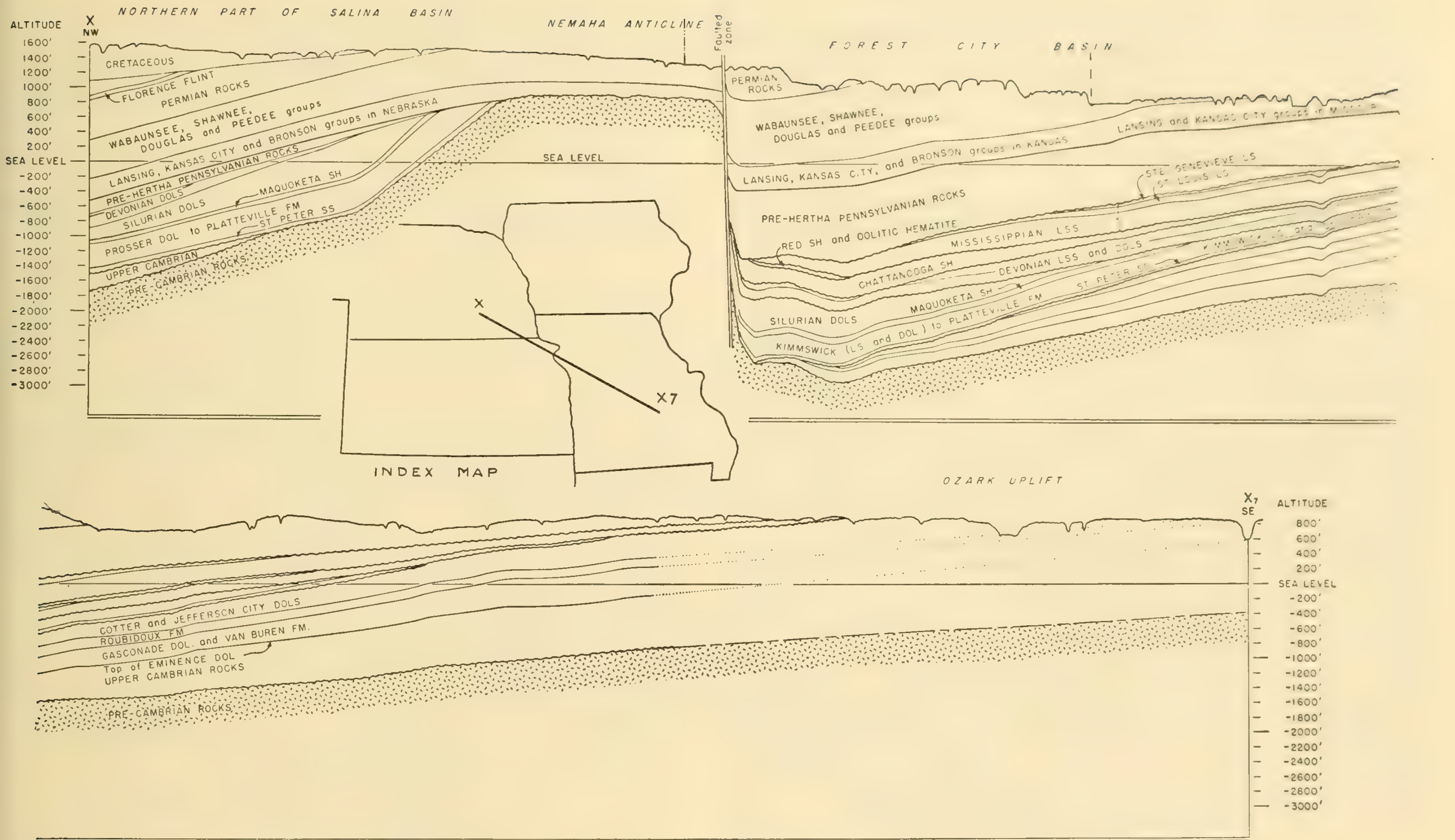


Fig. 5.14. Cross section of the Nemaha Range and Forest City basin. Reproduced from Wallace Lee *et al.*, 1946. Note the several unconformities.

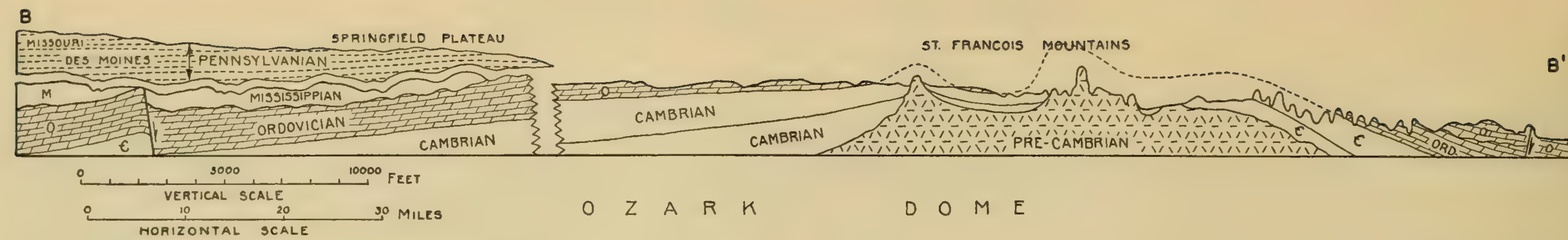


Fig. 5.15. Taken from *Geologic Map of Missouri* (1939). Section runs from Mt. Vernon to Perryville, Mo. The wedge of Pennsylvanian strata at the left is added to show the Mississippian and Pennsylvanian unconformity as it occurs about 50 miles north of the line of cross section. Section B–B' on map of Fig. 14.1.

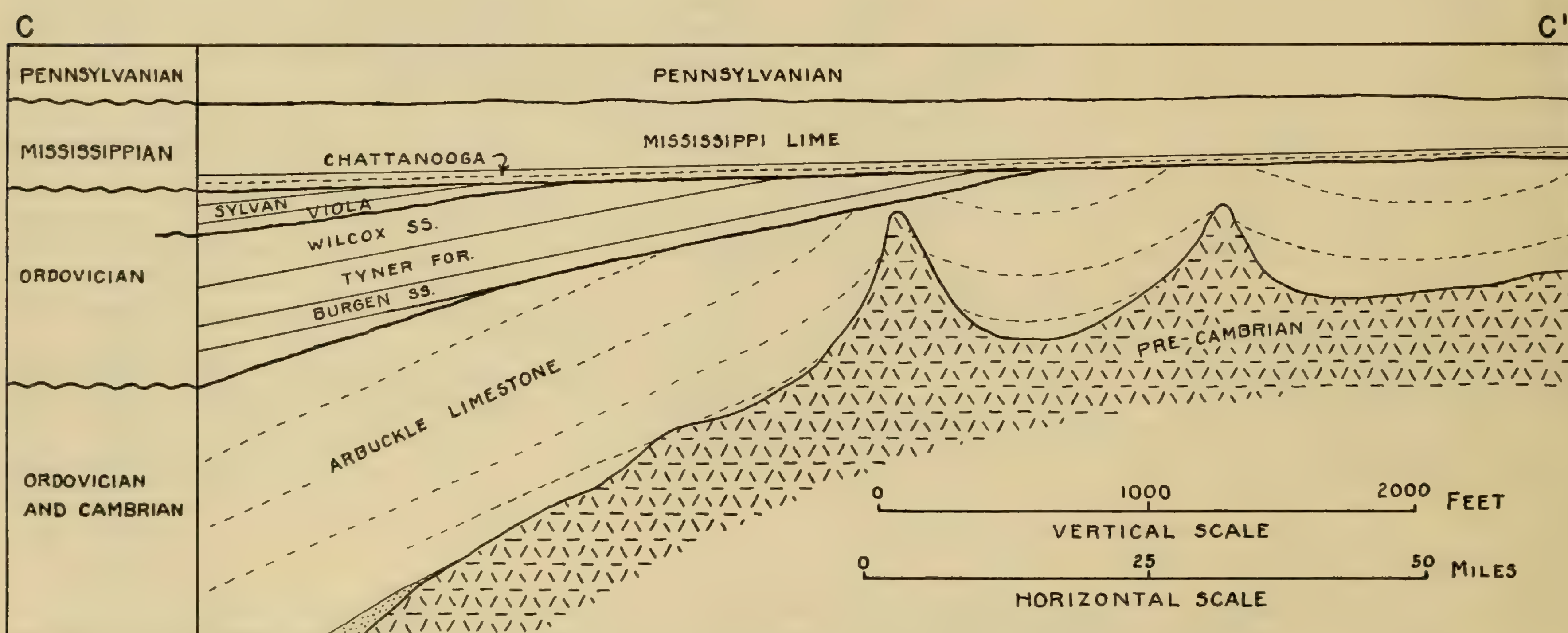


Fig. 5.16. Section across northeastern Oklahoma. Taken from White (1926, Pl. 1). Section C–C' on map of Fig. 14.1.

the Chautauqua arch, and the Ozark dome made up one continuous broad arch which left the Transcontinental Arch at right angles and veered eastward in southern Missouri.

The dome was uplifted again slightly in the late Mississippian (Plate 6). This time the movement was not in company with the Ellis and Chautauqua arches, but apparently with the Hunton arch to the southwest in Oklahoma (Dott, 1934). The great Pennsylvanian transgression nearly, if not entirely, covered the dome (Plate 6), and no recurrences of uplift during the Pennsylvanian or Permian have been described. The Devonian and Mississippian uplifts left the dome wrinkled with very gentle narrow folds that trend in a northwest direction.

The Arkansas Valley lies south of the Ozark dome and north of the complexly folded and thrust-faulted Ouachita Mountains. It is a structural basin as well as valley, and will be described in Chapter 14 under the heading, "Ouachita System."

Cambridge Arch

A number of wells which have penetrated "granite" have been drilled through the Pennsylvanian formations in a line running northwesterly across Nebraska (Ballard, 1942). Isopach maps along this row of wells suggest that the central Kansas arch, well known from many wells, continues northwestward to the Black Hills and beyond to the southeastern corner of Montana. The arch across Nebraska is known as the Cambridge arch (Plate 7). Geologic contacts determined from both surface and subsurface data, however, do not reveal the arch, because it lies mostly within the Precambrian rocks of the larger Transcontinental Arch (Plates 4 and 5). No wells have yet been drilled to the Precambrian northeast of the Cambridge-central Kansas arch, and therefore the boundaries of the pre-Pennsylvanian formations along the Transcontinental Arch may have to be shifted considerably at a later date.

NORTHWESTERN INTERIOR BASINS AND ARCHES

Williston and Alberta Basins

The Williston basin was first thought of as a gentle Tertiary downwarp in western North Dakota and eastern Montana, and was named after the town of Williston, N.D., on the Missouri River. Cretaceous strata were known to underlie the Tertiary and these to cover Paleozoic rocks of the extensive region of South and North Dakota, Montana, southwestern Manitoba, and southern Saskatchewan. With the discovery of commercial oil in 1951 in North Dakota, the term Williston basin became applied to the Paleozoic strata more particularly than to the Tertiary or Mesozoic, and with the drilling of many holes the distribution of formations and systems has become well known. Isopach maps of the several systems important in the Williston basin are shown in Figs. 5.17, 5.18, and 5.19.

A vast region in Alberta, western Saskatchewan, northeastern British Columbia, and the Mackenzie area of the Northwest Territories is a continuation of the Paleozoic sequence of the Williston basin, and the accompanying maps show the close relationship of the geology of the two regions, although they are generally treated separately in oil field parlance. The term "Alberta shelf" has been applied to the Paleozoic sedimentary province under the Great Plains of western Canada, because it is a shallowing shelf region to the Cordilleran geosyncline or Alberta trough on the west for most of the systems (Webb, 1954). It is also commonly referred to as the Alberta basin as a region for oil exploration and structurally as the Alberta syncline. During the Devonian period a broad basin did develop (see Fig. 5.17C), but otherwise the region can more properly be called a shelf. The syncline developed as the result of Cretaceous and Tertiary subsidence, mountain building on the west, and sedimentation, but the synclinal axis is not reflected under the Great Plains in the thicknesses of any of the pre-Cretaceous systems.

The Cambrian strata are dominantly clastic with a sandstone generally at the base and a sequence of green and maroon shales and light gray calcareous siltstones and fine-grained sandstones above. These beds were deposited unconformably on a Precambrian terrane as the seas invaded the shield region from the west and southwest (Fig. 5.17A).

The Ordovician beds are extensive under the Williston basin but generally absent on the Alberta plains. The outcrops in Manitoba contain a 50- to 100-foot basal, white quartz sandstone with interbedded shales

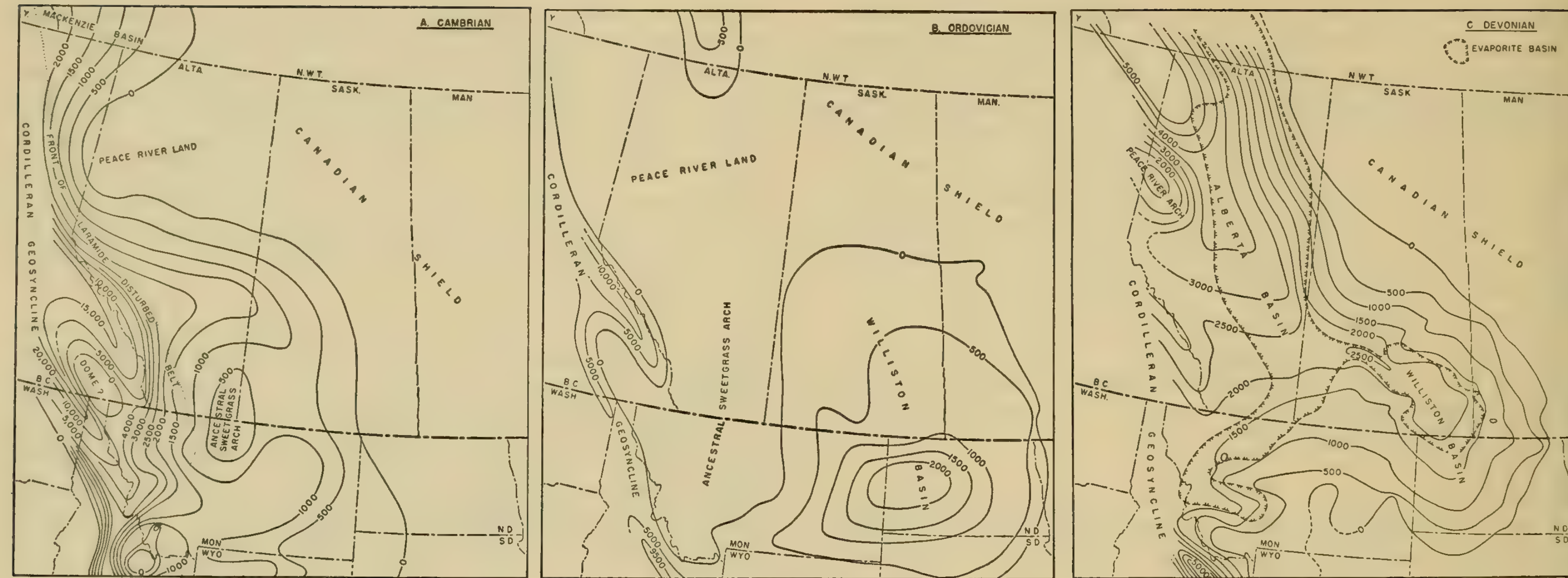


Fig. 5.17. Thickness map of Williston and Alberta basins: Cambrian, Ordovician and Devonian. Cambrian, after Webb (1954) and Sloss (1950); Ordovician, after Webb (1954) and Sloss (1950); Devonian with evaporite region, after Webb (1954), Sloss (1950), and Baillie (1955).

and then a sequence of 400 feet of limestones and dolomites. The carbonates are the chief rocks encountered in wells; the basal clastics appear to wedge out to the northwest (Webb, 1954).

The Silurian is represented in east-central Alberta Plains by an evaporite sequence and is generally included with beds which may be Middle Devonian. The Silurian and Middle (?) Devonian beds are the Elk Point formation of the stratigraphic chart, Figs. 5.20 and 5.21, and contain a composite salt thickness of 1200 feet in 1700 feet of beds. The Silurian is present in Manitoba, North Dakota, much of Saskatchewan and northern Montana, but with the Ordovician, is absent in the Sweetgrass arch region. It consists of light yellowish gray and yellowish orange, finely crystalline to dense dolomite (the Interlake group).

The Upper Devonian strata in western Canada are much more widespread than the Middle, and the original extent was still greater. Post-Paleozoic erosion has removed the beds over considerable areas. The Upper Devonian is characterized by thick deposits of limestones, dolomites, shales, and evaporites. It marks a time of limestone reef growth on widespread banks with numerous local bioherm and biostrom deposits and abrupt facies, changes, all holding large oil reserves.

The Devonian succession of the Williston basin is shown on the chart of Fig. 5.21, and its distribution in Fig. 5.17C. It is divided into four major lithologic units, which in ascending order are, Elk Point group, Manitoba group, Saskatchewan group, and Qu'Appele group. The lower three are chiefly carbonates but the upper is composed of red shales and siltstones. An extensive evaporite sequence occurs in the lower Elk Point group and also in the Manitoba group. In north-central Montana

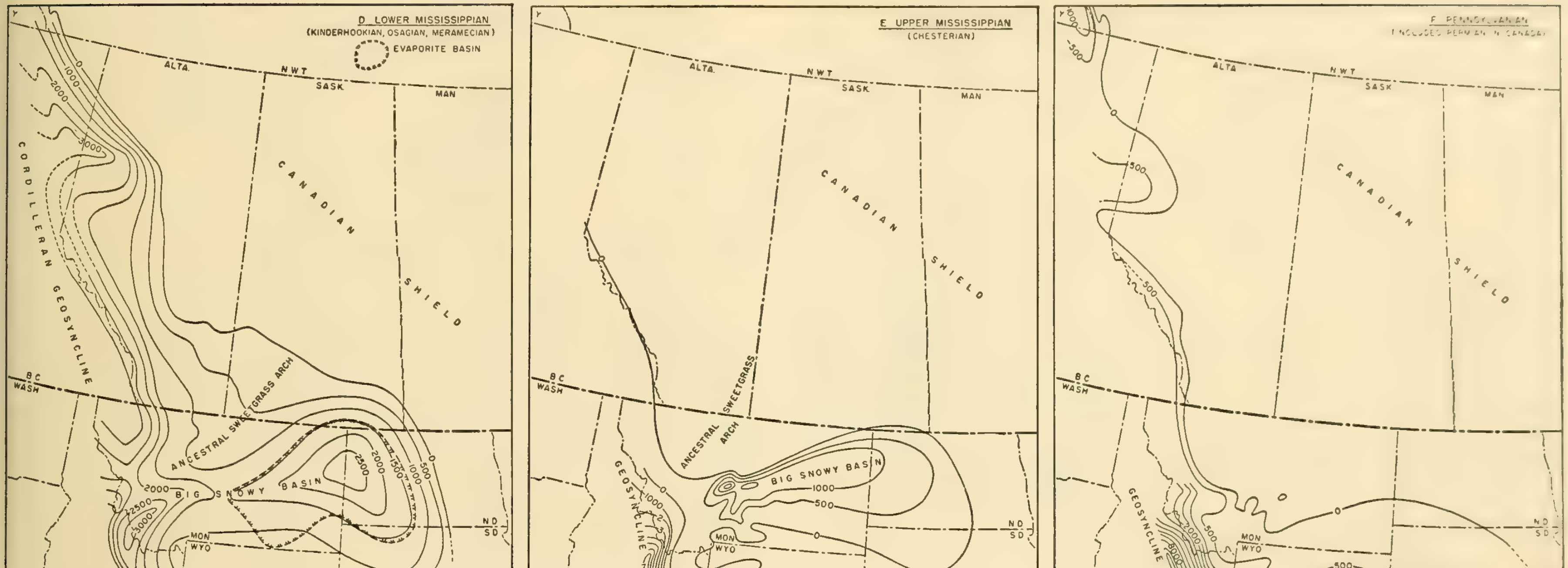


Fig. 5.18. Thickness map of the Williston and Alberta basins: Lower and Upper Mississippian and Pennsylvanian. Lower Mississippian (Kinderhookian, Osagian, and Meramecian series), after Webb (1954) and Sloss (1950); Upper Mississippian (Chesterian), after Sloss (1950); Pennsylvanian, includes Permian in Canada, after Webb (1954) and Sloss (1950).

a third evaporite sequence occurs at a still higher stratigraphic position, in the top of the Jefferson.

The Mississippian beds which rest on an erosion surface on the Devonian are marked at the base by black shale in the Williston basin. The Mississippian is more restricted in the Alberta region, but the beds possibly extended east at the time of deposition as far as the present margin of the Canadian Shield. The beds in Alberta start with a lower dark gray calcareous shale or dark brown-gray argillaceous limestone with fine-grained sandstone beds in the south. The upper beds are buff crystalline to dense limestones. The succession in the Williston basin beginning with the Kinderhookian and Osagian strata is largely limestone. These beds make up the Lodgepole and Mission Canyon formations. The Meramecian is dominated by dolomites which compose the Charles formation. See Fig. 5.22. The Charles contains considerable thicknesses of evaporites. See map, Fig. 5.18D.

The Upper Mississippian or Chester beds lie in an east-west basin through central Montana, called the Big Snowy. The eastern part of this basin, however, is in the general region of the Williston basin and hence it is considered part of the Williston. The strata are dominantly clastic in contrast to the chemical precipitates of the Lower Mississippian and compose the Kibbey, Otter, and Heath formations. Also part of the overlying Amsden formation is Chester in age.

The Alberta shelf region was emergent and suffered long-continued erosion during the Pennsylvanian. In the front ranges of the Rockies, however, a thin sequence of sandy dolomites and quartzitic and cherty sandstones are Pennsylvanian and Permian in age, and are known as

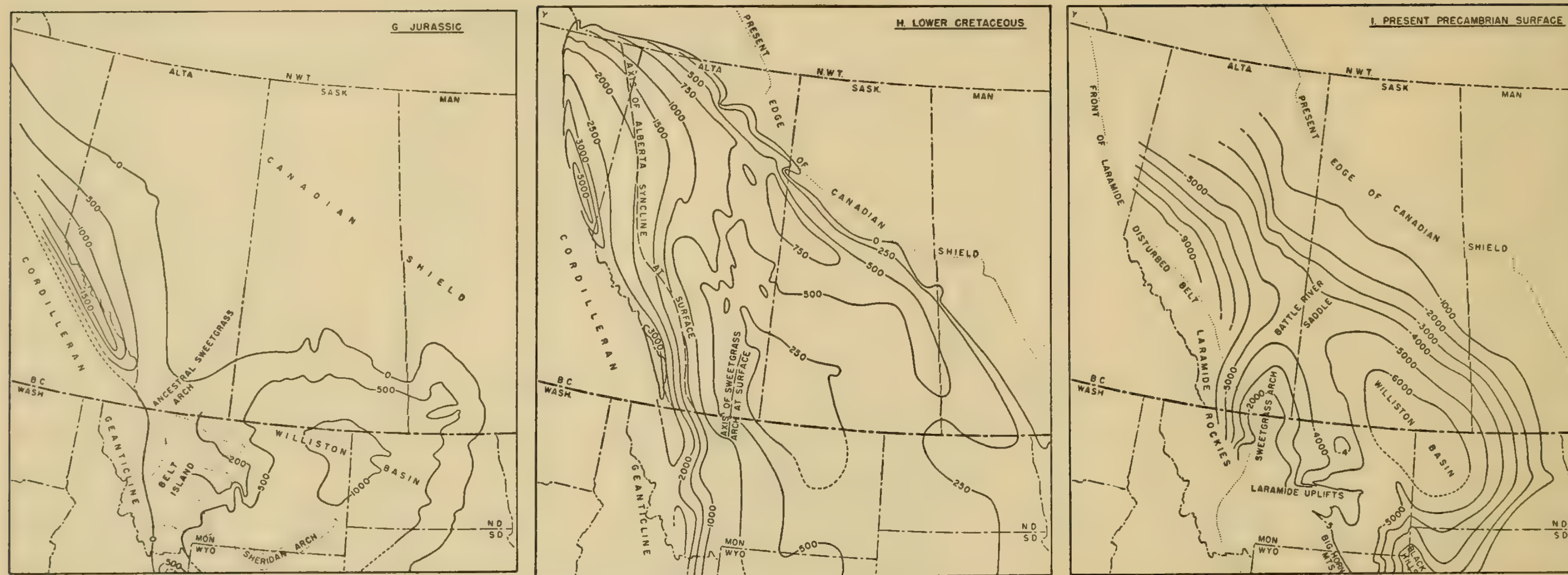


Fig. 5.19. Thickness maps of Williston and Alberta basins: Jurassic and Lower Cretaceous. Also contour map on Precambrian surface. Jurassic, after Webb (1954), Francis (1957), and Peterson (1957); Lower Cretaceous, after Webb (1954) and Reeside (1944); Precambrian surface, from *Tectonic Map of Canada* (1954) and Moss (1936).

the Rocky Mountain formation. Farther north in adjacent parts of Yukon and Northwest Territories equivalent sandstones with a chert member at the top attain a maximum thickness of 1200 feet. The erosion surface on the Mississippian in the Peace River region has local sharp relief, and beds believed to be Pennsylvanian and Permian cover the surface and range up to 500 feet thick.

In the Montana and South Dakota area (see map, Fig. 5.18F) clastics predominate over non-clastics, and clean quartzose sandstones are the rule, making up the Quadrant sandstone in central and western Montana and the Tensleep sandstone over the Wyoming shelf. In the southern part of the Williston basin a wedge of Pennsylvanian is preserved, and consists of dolomite interbedded with sandstone, red shale, and evaporites.

Triassic time was marked by widespread emergence, but in the Peace River Country a thick sequence of marine clastics, impure limestones and anhydrite accumulated. Thicknesses up to 3000 have been measured in the adjacent Rockies.

A group of red beds has been charted across part of the Williston basin by Ziegler (1956). The beds lie between the Permian Minnekahta limestone and the Piper beds of the Jurassic. See map, Fig. 5.22. A lower shale and siltstone unit is thought to correlate with the Spearfish red beds of the Black Hills which are Triassic, and three overlying units, a salt, a siltstone and sandstone, and an upper salt are thought to be lower Jurassic but may also be Triassic.

The Jurassic beds in Alberta have about the same distribution as the Triassic except for a wider transgression in the southern Foothill belt

			Front Range and Foothills	Central and Southern Plains	Southeastern Plains
CENOZOIC	TERTIARY	Pliocene Miocene Oligocene Eocene Paleocene	Paskapoo	Cypress Hills Swift Current Ravenscrag	Wood Mountain Turtle Mountain
MESOZOIC	CRETACEOUS	Upper	Edmonton Belly River Wapiabi Bighorn (Cardium)	Edmonton Bearpaw Pale Beds Foremost Pakowki Milk River Lea Park Alberta (Colorado)	Boissevain Riding Mountain Pembina Boyne Morden Favel
		Lower	Blackstone Blairmore Upper Kootenay	Blackleaf (Viking) member (Bow Id.) Blairmore	Ashville Swan River
	JURASSIC	Upper Middle Lower	Lower Kootenay Fernie	Ellis group	Morrison ? Sundance Gypsum Springs (Amaranth ?)
	TRIASSIC	Upper Middle Lower	Schooler Creek Spray River		
PALEOZOIC	PERMO-PENNSYL		Rocky Mountain		
	MISSISSIPPIAN	Upper Lower	Rundle Banff	Madison group	Charles Madison Kinderhook
	DEVONIAN	Upper Middle Lower	Exshaw Palliser Fairholme ? Ghost River ?	Exshaw Wabamun (Potlatch) Winterburn (Jefferson) Woodbend Beaverhill (Waterways) Elk Point (upper) (part)	Exshaw Lyleton Jefferson Manitoban Winnipegosan Elm Point
	SILURIAN	Upper Middle Lower	[Ghost River ?]	Elk Point ? (lower) (part)	Ashern Interlake group Stonewall
	ORDOVICIAN	Upper Middle Lower			Stony Mountain Red River Winnipeg
	CAMBRIAN	Upper Middle Lower	Cathedral ?	Upper Cambrian	
	PRECAMBRIAN		Late Proterozoic sediments	Chiefly Archean Intrusives and metamorphics	

Fig. 5.20. Generalized correlation chart of western Canada basin, southern part, after Webb, 1954.

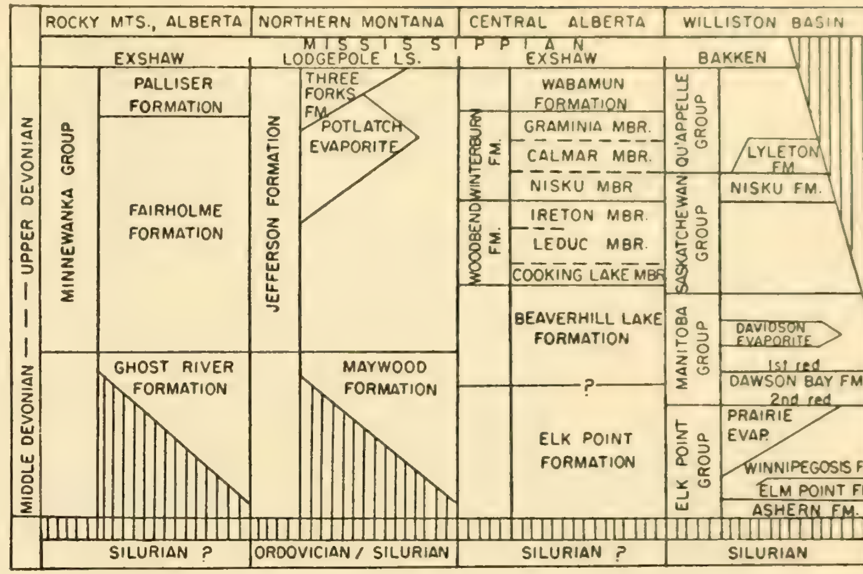


Fig. 5.21. Devonian correlation chart of the Williston and Alberta basins. Reproduced from Baillie, 1955.

near Calgary. There, a fairly thick succession representing Lower, Middle, and Upper Jurassic occurs. Over the Sweetgrass arch (Fig. 5.19G) only a thin marine sequence of shales and sandstones of Middle and Upper Jurassic beds is present. These rest on an irregularly eroded surface of the Mississippian. Peterson (1957) traces the depositional history of western Montana, and for the intermittently positive area where thinning and overlap occurred he uses the term Belt Island, but explains that it was rarely emergent and then only in small areas. It had been emergent in early Jurassic time and probably furnished some of the clastic material for the adjacent Middle Jurassic formations. See chart, Fig. 5.23. Another area that tended toward shoal conditions during parts of mid- and late Jurassic time, although not emergent, was the Sheridan arch. Middle and Upper Jurassic beds are widespread over the Williston basin and define it in about the position of the older Mississippian basin but centered somewhat south of the Devonian basin.

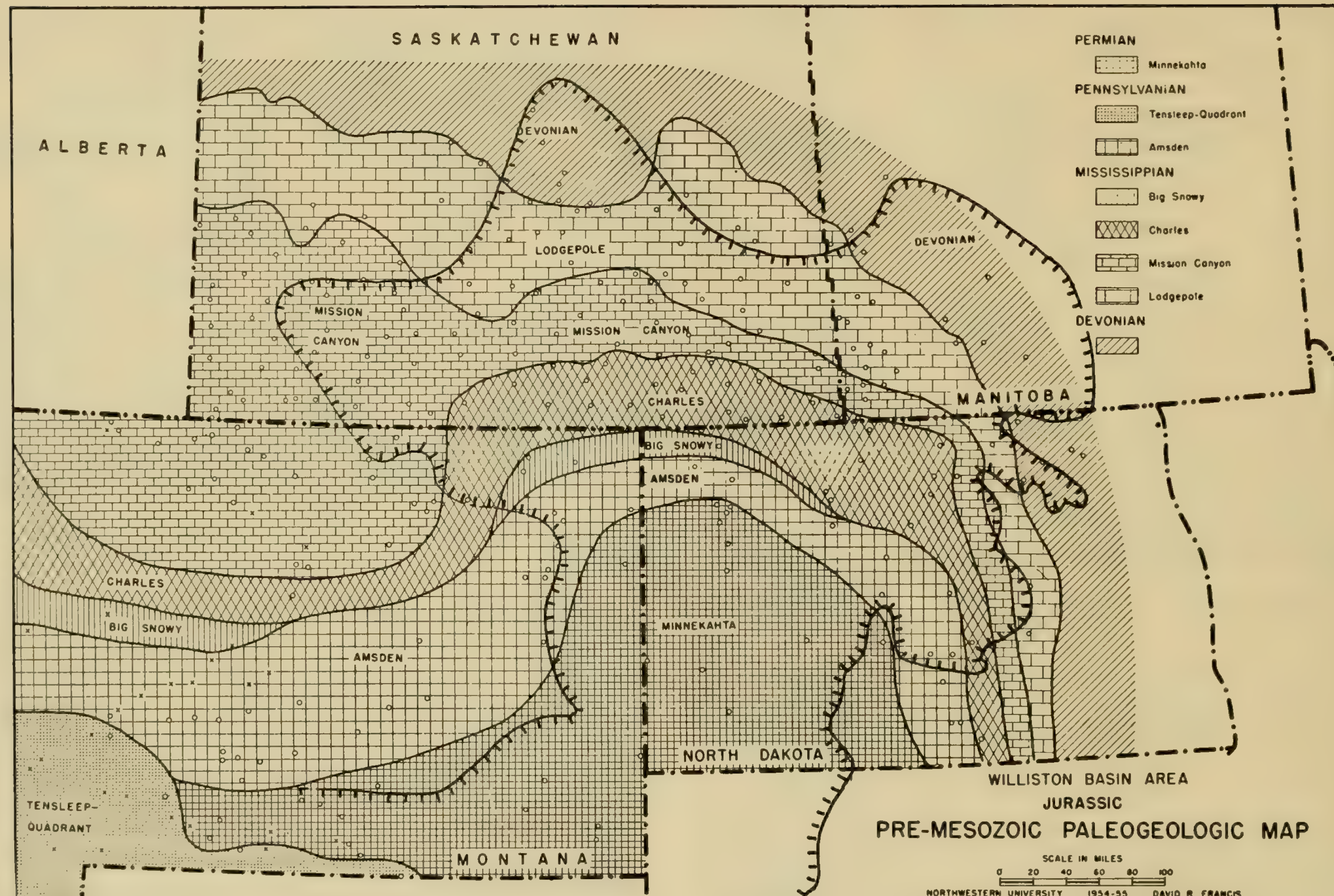


Fig 5.22. Distribution of formational outcrops before Mesozoic strata were deposited unconformably over the Paleozoic strata. The hachured line indicates extent of Triassic (?) red bed deposition (from Ziegler, 1956). Jurassic sediments spread over almost entire area. Map reproduced from Francis (1956).

By Jurassic time the rise of the Cordilleran geanticline had become extensive (see Plate 10 of Chapter 3), and considerable sediment was shed from it eastward to the subsiding areas of accumulation. Part of the geanticline became engrossed in major mountain building in Early Cretaceous time, and this, The Nevadan orogeny, resulted, in British Columbia, in the uplifting, disruption, and widespread intrusion of the sedimentary rocks of the Cordilleran geosyncline. A new restricted trough or longitudinal basin formed, as shown in Figs. 5.19 and 5.20, in about the position of the present Canadian Rockies. The Nevadan Orogeny engrossed the Selkirk Range on the west as well as a vast region westward to the continental margin. The earliest Lower Cretaceous sediments deposited were a thick coal-bearing series, the Kootenay formation, and then after a brief erosion interval clastics of the Blairmore formation spread eastward over the Kootenay and extensively over the Alberta shelf region. See Fig. 5.19H. The coarse clastics along the foothills and front ranges of the Rocky Mountains and maximum thickness there indicate that the rise of the mountain belt on the west was rapid, and that it was suffering active erosion.

The distribution of Upper Cretaceous sediments is about that of the Lower Cretaceous and follows about the same pattern of thickening westward into the trough. The Upper Cretaceous are much thicker than the lower Cretaceous in the Williston basin and attain thicknesses of 4000 feet in eastern Montana and the western part of the Dakotas. The Upper Cretaceous beds reflect the growth of the later Laramide Rockies and become involved themselves in deformation. They, with a central blanket of Tertiary beds, have been deposited and gently folded adjacent to the major belt of mountain building on the west to form the Alberta syncline.

A contour map of the pre-Paleozoic surface reflects the summation of all subsidences and uplifts in the Alberta–Williston region, and it will be seen (Fig. 5.19I) that the center of the Williston basin is about at the international boundary and the North Dakota–Montana line. All told, it now holds over 7000 feet of sediment. Its position and extent are somewhat modified by the central Montana and Black Hills uplifts of Late Cretaceous age. The Sweetgrass arch is a strong element of 4000 feet

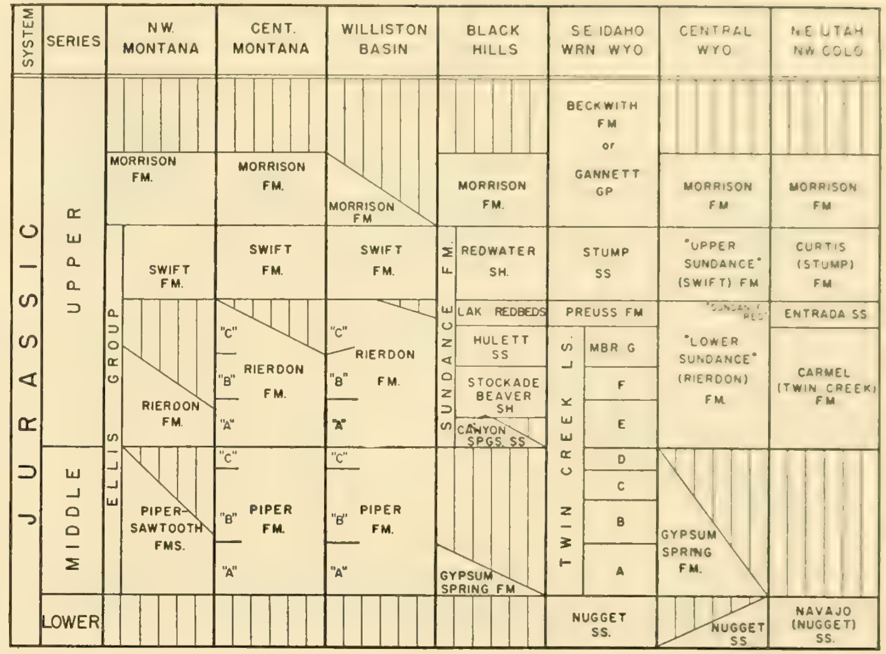
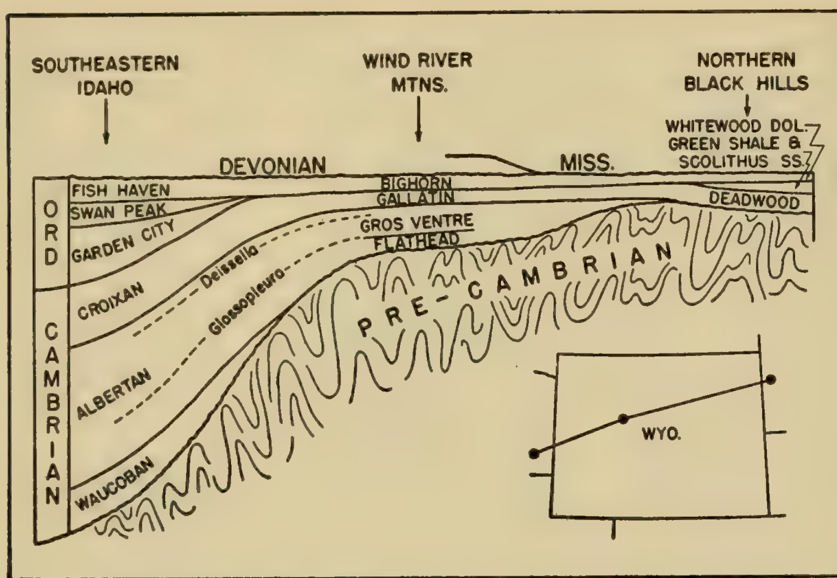


Fig. 5.23. Jurassic correlation chart of the Williston basin and adjacent areas. Reproduced from Peterson, 1957.

relief. The Alberta basin centers between Peace River and Edmonton, and contains there in front of the disturbed belt over 10,000 feet of sediments. Within the disturbed belt the thickness is much greater, and had the Precambrian surface not been broken and deformed in the Nevadan and Laramide orogenies it would lie very deep, indeed.

Utah-Wyoming Shelf

The Williston basin and its relation to the Alberta shelf has already been described. Southward through central and eastern Wyoming and the Colorado Plateau of Colorado and Utah relatively thin layers of Cambrian, Ordovician, Devonian, Mississippian, Pennsylvanian, and Permian strata occur. They represent the transition from the geosyncline on the west to the Transcontinental Arch on the east. The influence of the Ancestral Rockies and other land movements in Carboniferous time



Stratigraphic diagram showing the relations of Cambrian and Ordovician rocks between southeastern Idaho and the northern Black Hills.

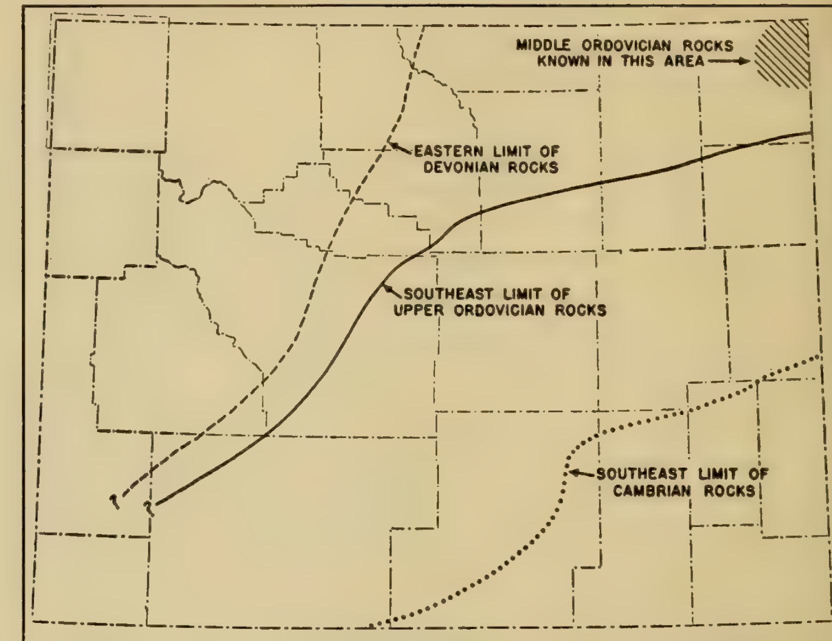


Fig. 5.24. Relation of shelf in Wyoming to Transcontinental Arch and Cordilleran geosyncline. Reproduced from Thomas, 1949.

on the sites of deposition is shown on Plate 7, and Fig. 6.7. Figure 5.24 is a cross section to illustrate the shelf and its relation to the Transcontinental Arch and the Cordilleran geosyncline.

The formations of the Wyoming and Montana part of the Utah–Wyoming shelf differ somewhat from those of the Utah and Arizona part. The formations have been the object of numerous stratigraphic studies because of their importance as oil and gas producers. See correlation charts listed in Chapter 1.

6.

PALEOZOIC CORDILLERAN GEOSYNCLINE

DIVISIONS AND THEIR CHARACTERISTICS

Schuchert is probably more responsible than anyone else for the use of the expression Cordilleran geosyncline in describing the basins of accumulation of sediments along the western margin of the continent. He also used the term Rocky Mountain geosyncline. During the Mesozoic, his "Cordilleran intermontane geanticline" split the overall broad and irregular basins into two longitudinal divisions, but before the geanticline became pronounced, the divisions were already evident by the nature of their sediments, the western being an eugeosynclinal assemblage and the eastern a miogeosynclinal. The eugeosyncline extended from mid-Nevada to the Pacific Coast, and the miogeosyncline from mid-Nevada to central Utah (Fig. 6.1). The miogeosyncline is much better known than the eugeosyncline. The basins of sedimentation and geography shifted somewhat from one period to another, but the broad overall relations remained fairly constant. The change from the thick sedimentary sequence of the miogeosyncline to the thin sediments of the shelf has been called the Wasatch line (Kay, 1951), and for all Paleozoic periods except Silurian the change is fairly abrupt and in much the same position. The broad divisions as outlined were probably first recognized by Stille (1941) and later elaborated on by Kay (1942, 1951, 1960) and Eardley (1947).

The eugeosyncline probably sank more and received a greater thickness of sediments than the miogeosyncline, but the extent of sediments in both was great. The major difference lies in the character of the sediments. The eugeosyncline received a dominant amount of volcanic material and graywacke, whereas the miogeosyncline was filled with sandstones, quartzites, shales, limestones, and dolomites. The volcanic material in the eugeosyncline is in several forms: flows, volcanic conglomerates, and various pyroclastics. The volcanics and graywackes occur in every stratigraphic system from Upper Cambrian to Cretaceous. The Permian especially was a time of excessive volcanism, and the volcanics of that period have been traced from California and western Nevada to Alaska (Wheeler, 1939; White, 1959). In the Humboldt Range of northwestern Nevada, over 10,000 feet of Permian strata, largely volcanic, have been identified.

Roberts *et al.* (1958) estimate that the miogeosynclinal strata in eastern Nevada and western Utah above the thick basal quartzite of the Cambrian consist of 60 per cent limestone, 30 percent dolomite, 8 percent shale, and 2 percent quartzite. They estimate that the eugeosynclinal strata, on the other hand, in the Sonoma Range and vicinity consist of 20–40 percent shale, 10–30 percent sandstone, graywacke, and quartzite, up to 30 percent of chert, with shale partings, and up to 30 percent of volcanic and pyroclastic rocks.

> The units are characteristically lenticular, and thin or thicken abruptly parallel with and normal to the geosynclinal trend. Limestone, generally shaly or sandy, locally forms thin, discontinuous layers. The shale units are commonly

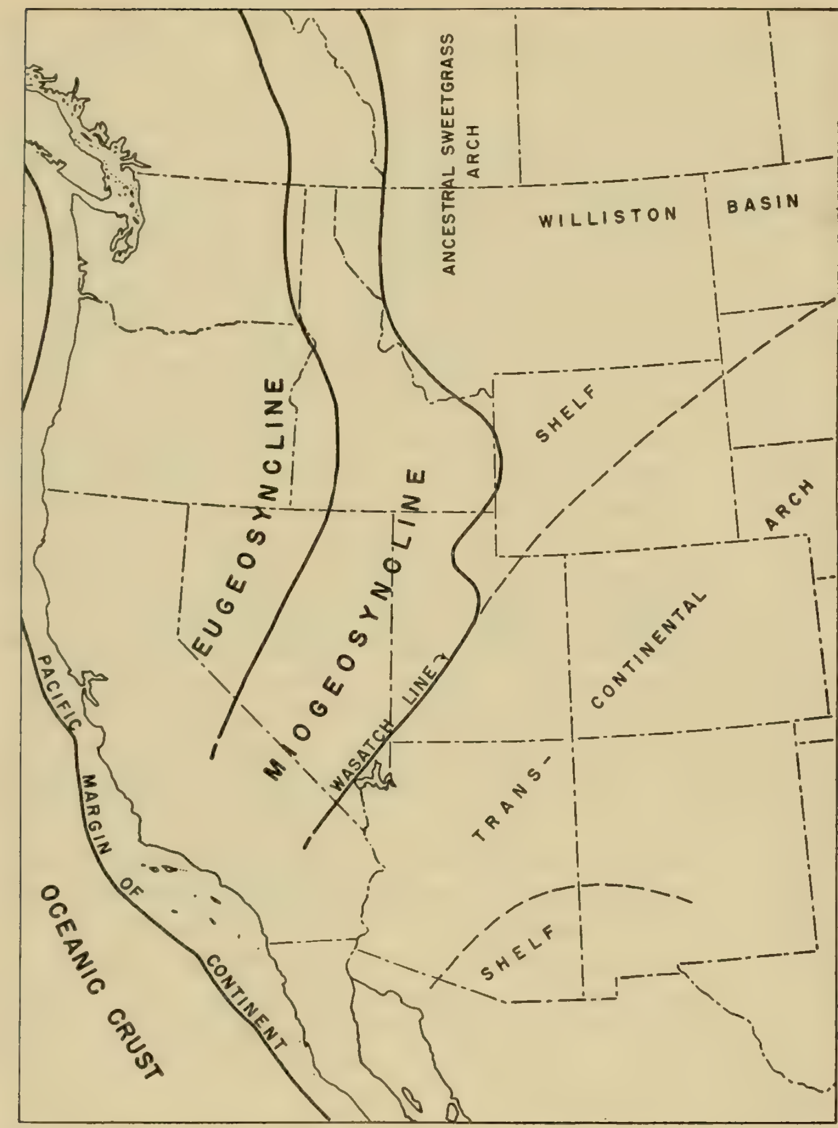


Fig. 6.1. Major Paleozoic tectonic elements of western United States. The eugeosynclinal bondary was farther east than shown in Permian time. The Wasatch line through southern Nevada has been called the Las Vegas line (Welch, 1959).

sandy and few are calcareous. The quartzites are generally nearly pure, but the sandstones are either graywackes or feldspathic sandstones. The chert units, partly of volcanic derivation, range from a few inches to several hundred feet thick; individual chert layers are lenticular and range from a fraction of an inch to 3 feet. They are separated by shaly partings which are also lenticular; laterally, chert units grade into siliceous shale units with subordinate chert. The volcanic rocks are largely andesitic or basaltic pillow lavas and pyroclastics that accumulated mainly in a marine environment; most are highly albitic. Siliceous pyroclastic rocks locally form thick sections. The volcanic rocks are highly lenticular, and probably formed around many source centers (Roberts *et al.*, 1958).

Another characteristic of the sediments of the eugeosyncline is their metamorphism. The thick sequences, especially in the Sierra Nevada, Klamath Mountains, western British Columbia, and southeastern Alaska, are made up of phyllites; slates; argillites; quartz, chlorite, hornblende, and calcareous schists; hornblende gneiss; recrystallized chert; marble; meta-conglomerate; meta-andesite; and various metamorphosed pyroclastics. Still another characteristic is the presence of great intrusive bodies of later age, and the metamorphism of the sediments about the intrusions.

The sediments of the miogeosyncline, on the other hand, are not metamorphosed. Many of the sands are cemented with silica and termed quartzite, but little dynamic metamorphism incident to Paleozoic, Mesozoic, or Tertiary orogeny has occurred.

The medial belt in central Nevada contains transitional types of the two environments, and became not only a geanticline but a belt of orogeny in late Devonian time. The western eugeosynclinal strata were thrust many miles eastward to rest on miogeosynclinal strata of strikingly different lithology.

BASINS AND UPLIFTS OF THE WESTERN UNITED STATES AND SOUTHERN BRITISH COLUMBIA

Cambrian Basins

The miogeosyncline is noted for its Cambrian sections (Fig. 6.2). At one locality in southern Nevada and California 17,000 feet of Lower, Middle, and Upper Cambrian beds have been measured.

The oldest Cambrian rocks over much of eastern and southern Nevada and southwestern Utah is the Prospect Mountain quartzite, which may be over 5000 feet thick in places. The Osgood Mountain quartzite in north-central Nevada, the equivalent of the Prospect Mountain, may be as much as 10,000 feet thick. Overlying the quartzite are shale, dolomite, and limestone formations of uniform and wide occurrence. Stratigraphic sections from the eugeosyncline to the miogeosyncline of north-central Nevada are shown in Fig. 6.9, and of the miogeosyncline of western and northern Utah in Figs. 6.9 and 6.10.

In southeastern British Columbia is another succession of Cambrian strata which totals about 10,000 feet in maximum thickness. From the Burgess shale of this succession Wolcott took an amazing assortment of fossils and greatly enriched our knowledge of life at the beginning of Paleozoic time. Lower Cambrian beds are absent at the international boundary, but further north in the Mount Robson vicinity they are present and consist of 3900 feet of quartzitic sandstone, siliceous shale, and limestone. Upper Cambrian strata are restricted and consist mostly of limestone (Lord *et al.*, 1947).

Another thick Cambrian sequence is known in northeastern Washington where at least 12,000 feet of beds dated by fossils occur. The Gypsy quartzite lies at the base; over this is the Maitlen phyllite, and over this the Metaline limestone (Park and Cannon, 1938; Campbell, 1947). The assemblage is miogeosynclinal in aspect and contains elements of the same fauna as the miogeosyncline of western Utah and eastern Nevada (Wm. Lee Stokes, personal communication).

Representative of the eugeosynclinal assemblage in Cambrian time is the Scott Canyon formation in Battle Mountain. It is composed of greenstone, chert, and some shale, and is about 5000 feet thick (Roberts *et al.*, 1958).

Lower and Middle Cambrian sediments are just about entirely restricted to the geosyncline, but Upper Cambrian strata are spread widely over the Central Stable Region of the United States as far as Wisconsin and Ohio. Here they are overlapped by Ordovician sediments which extend to the north and northeast over the Precambrian rocks of the Canadian Shield.

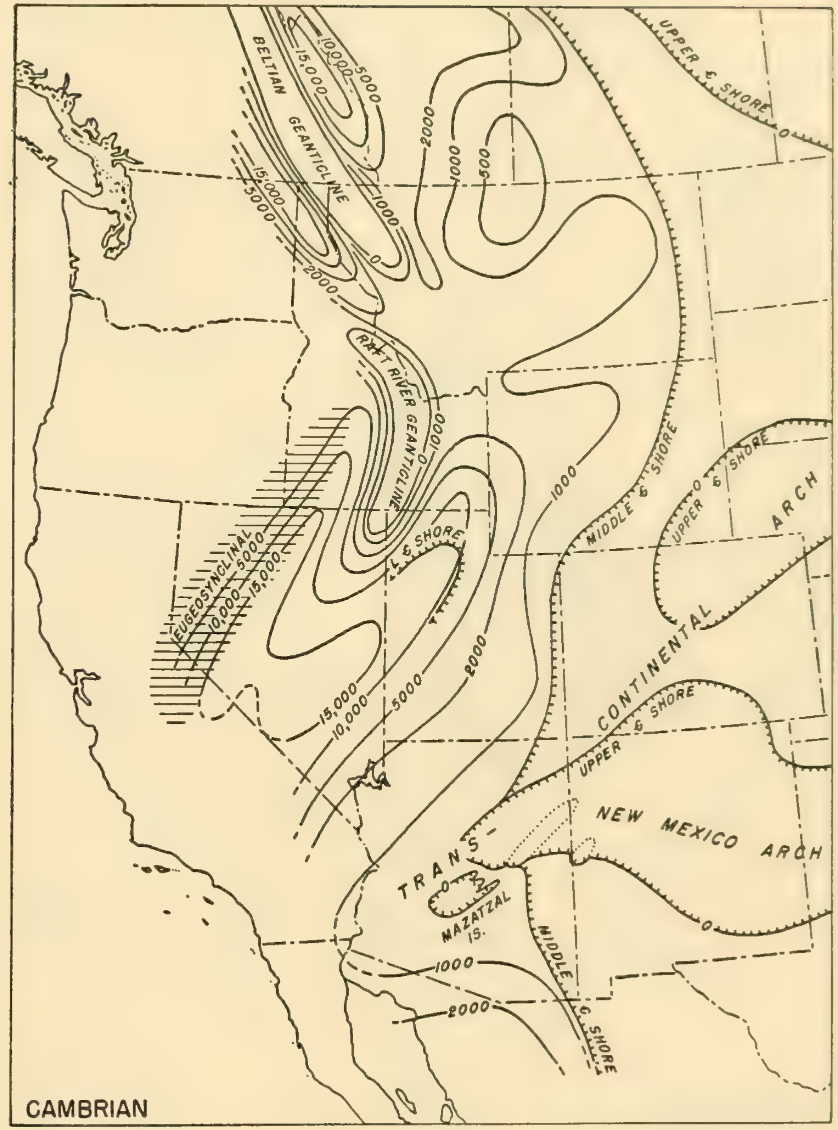


Fig. 6.2. Thickness and paleographic map of the Cambrian.

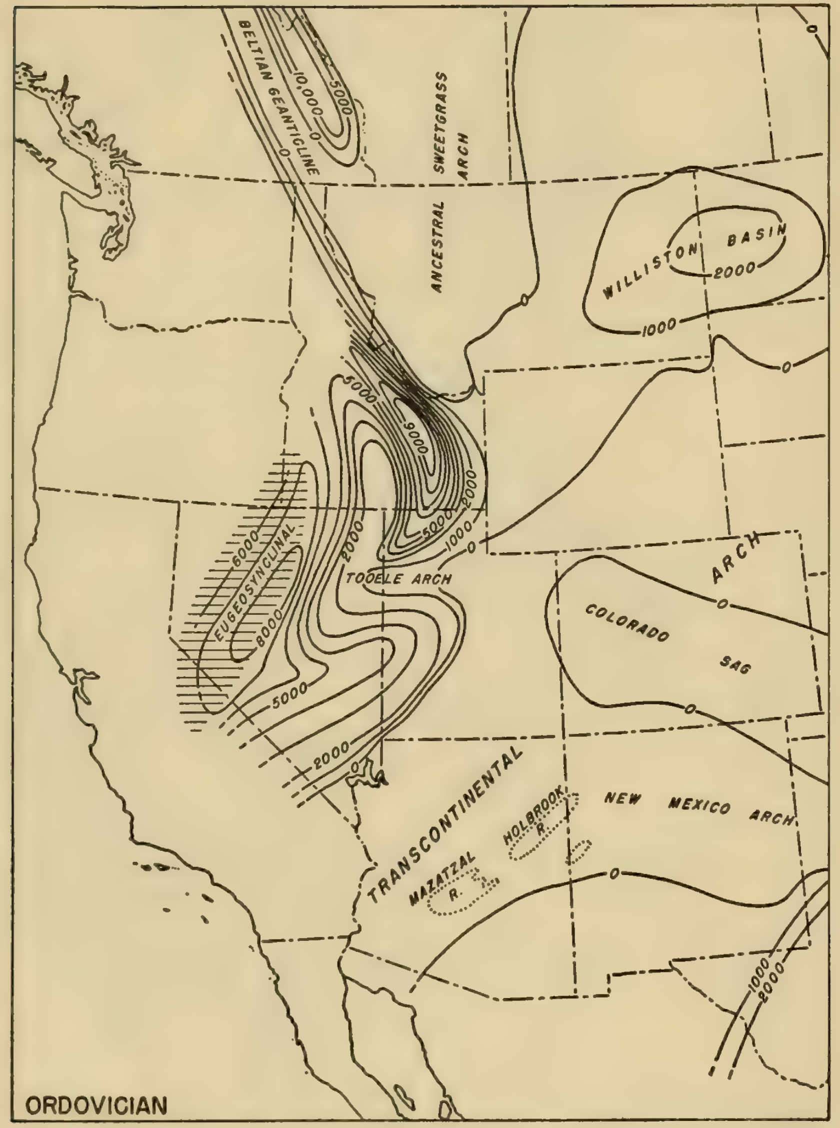


Fig. 6.3. Thickness and paleographic map of the Ordivician.

In Chapter 4 the Precambrian Mazatzal and Beltian orogenic belts have been described. Although the Beltian trough of sedimentation and later belt of orogeny marked the first tectonic development parallel with the present Pacific margin of the continent, the older Mazatzal belt seems to have made an impress on the Paleozoic geosynclinal basins. The Transcontinental Arch, which reflects the Mazatzal orogenic belt, borders directly on the arch in southeastern Utah, Arizona, and Colorado, and the two have the same trend to the southwest. See maps, Figs. 6.1 to 6.5

An uplift here called the Raft River geanticline, is identified in northwestern Utah (Stokes, 1952; Felix, 1956) and southwestern Montana (Scholten, 1957) on the south and north sides of the Snake River volcanic field respectively (see Fig. 6.11). Its extent northwestward cannot be told because of the cover of Tertiary volcanic rocks and the intrusion of the great Idaho batholiths, but in the interpretation rendered on Fig. 6.2 it appears as a geanticlinal uplift between the eugeosyncline basin in northern Nevada and the miogeosyncline of Utah. An unconformity in the Upper Cambrian detected in the South Stansbury Mountains (Rigby, (1958) with 700 feet of beds removed may be a lateral affect of the Raft River geanticline (see Fig. 6.10). The erosion surface lies beneath the Cole Canyon dolomite.

Still farther north in northwestern Montana, northern Idaho, and British Columbia is an extensive region of Precambrian strata, the Belt series, and this is here interpreted to have been a fairly persistent structural feature from Cambrian time on. Evidence cannot be sited for shoreline deposits and overlapping relations, but this is mostly due to the extensive batholithic intrusions and metamorphism. Early geologists considered the Beltian terrane the shore of an extensive, west-lying land which they called Cascadia, but later ones have considered the Paleozoic strata, beginning with Middle Cambrian, to have been deposited across and then eroded away incident to the emergence of the modern geanticline in Cretaceous and Tertiary times. Sloss (1950) however, suggests a small uplift there, and his interpretation is reflected on the maps of the Williston basin, Figs. 5.17, 5.18, and 5.19. The writer takes the view that it has been a significant feature from Cambrian time on (see Chapter 33).

No Cambrian or Ordovician fossils have been found in northern California, Oregon, and all Washington except the northeast corner. The lack of information about the western margin of the continent in Cambrian time, and in Ordovician as well, is disappointing. The oldest fossils yet discovered along the Pacific margin in the United States and British Columbia are Silurian. These have been found in the Klamath Mountains by Wells (1956). Three metamorphic series underly the fossiliferous Devonian strata there, according to Hinds (1939), and one or more of these might be Ordovician and Cambrian. See Fig. 6.3. In southeastern Alaska Buddington reports Ordovician fossils, but no Cambrian. In conclusion it may be assumed that the entire region west of central Nevada was eugeosynclinal from Ordovician time to the close of the Paleozoic.

Ordovician Basins

A broad Ordovician basin exists in western Utah and Nevada with miogeosynclinal type sediments in the eastern and eugeosynclinal type in the western part (see Fig. 6.3). The formations and their lithologies are shown in Fig. 6.9, which is a section across central Nevada and marks the change from the eugeosyncline to the miogeosyncline. The miogeosynclinal sediments of western Utah are reviewed by Hintze (1951) and summarized in the table of Fig. 6.12.

Another basin, which was narrower and completely miogeosynclinal in character (Fig. 6.12, Logan area), existed in southeastern Idaho and northern Utah. For a review of the stratigraphy see Ross (1953). In both basins the rocks are dominantly limestones and dolomites, but conspicuous quartzite formations exist in each. The Swan Peak quartzite of southeastern Idaho and northern Utah is about 500 feet thick, and the Eureka quartzite and the Swan Peak quartzite of western Utah and eastern Nevada are nearly 800 feet thick together. The Eureka quartzite, 537 feet thick at Ibex, Utah, overlies an 85-foot dolomite member, and this overlies the Swan Peak quartzite, 249 feet thick. The dolomite member wedges out east of Ibex, and there the upper quartzite rests directly on the lower. The absence or near absence of these sandstones together with a thinner Ordovician section in Utah southwest of Great Salt Lake indicates an uplift there which Webb (1958) has defined and named the

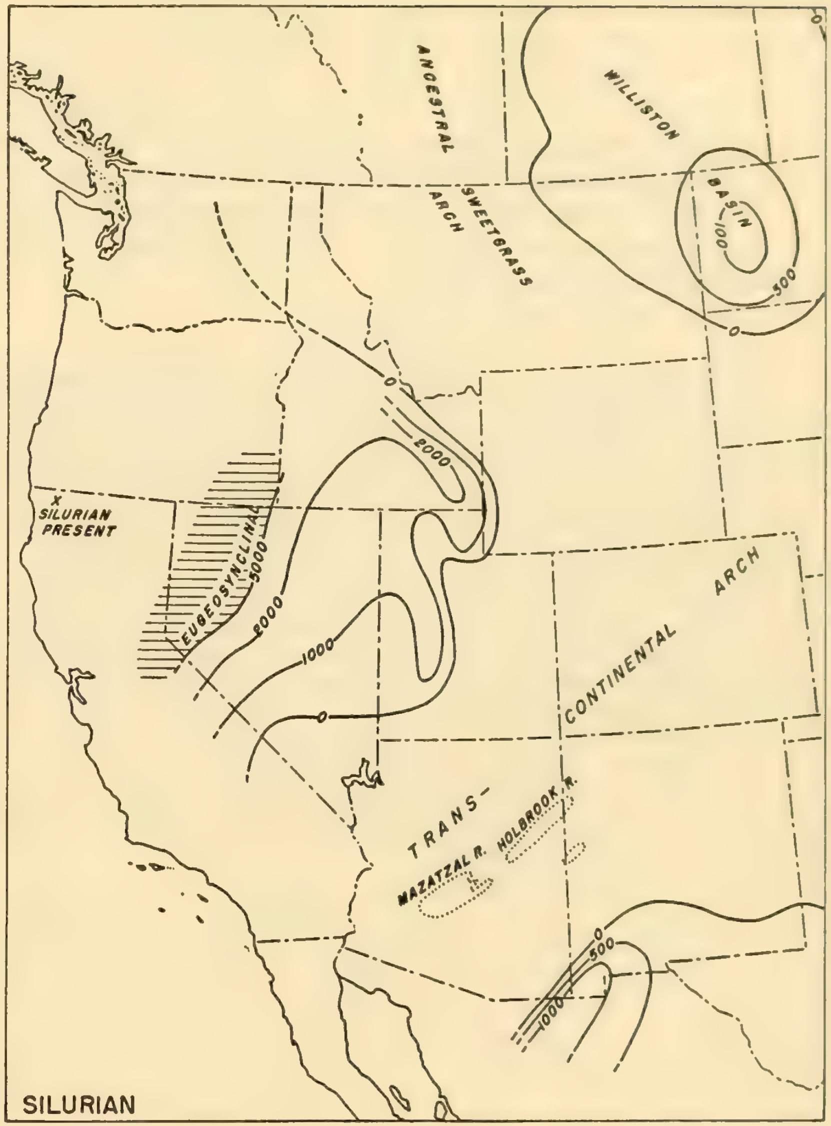


Fig. 6.4. Thickness and paleogeographic map of the Silurian.

Tooele arch. The arch and erosion is pre-Fish Haven (see Fig. 6.12).

A deep and evidently narrow trough of Ordovician sediments exists in the Canadian Rockies of western Alberta and eastern British Columbia. It is interpreted to lie east of the Beltian geanticline and to be separated by it from the basin of northeastern Washington containing the Ordovician Ledbetter slate, also of miogeosynclinal type. The Ordovician strata of the Canadian Rockies consist of 3000 to 7000 feet of limestone, shale, and slate beds with fossils representing a range from Lower to Upper in different places (Lord *et al.*, 1947).

According to Roberts *et al.* (1958):

Rocks of Ordovician age that belong to the western assemblage (eugeosyncline) are widely exposed throughout north-central Nevada. They underlie large areas in the Sulphur Spring Range, Roberts Mountains, Tuscarora Mountains, Cortez Mountains, northern Shoshone Range, Toyabe Range, Battle Mountain, and the Sonoma Range. So far as known they are allochthonous.

Merriam and Anderson (1942, p. 1694) used the name Vinini formation for rocks of Ordovician age of the western assemblage in the Roberts Mountains. They divided the formation into two units. The lower part of the Vinini, Early Ordovician in age, consists of quartzite, limestone, and calcareous sandstone, and silty and shaly sediments with minor amounts of andesitic lava flows and tuffs; perhaps the relatively abundant limestone here suggests an approach to the transitional assemblage. The upper part of the Vinini, of Middle Ordovician age, is composed of bedded chert and black organic shale, clearly of normal western lithologic type.

The most complete stratigraphic section of the Vinini formation thus far seen is in the Tuscarora Mountains, northern Eureka County, about 5 miles north of U.S. Highway 40. Strata of Early, Middle, and probably late Ordovician age are present; no detailed measurements were made, but it is estimated that the section is at least 7,000 feet thick.

In the Shoshone Range, Battle Mountain, and Sonoma Range the proportion of massive quartzite, chert, and volcanic material in the Ordovician rocks of the western assemblage is larger than in the Vinini formation. These rocks were named the Valmy formation in Battle Mountain (Roberts, 1949, 1951) where they have been subdivided into two members. The lower part of the Valmy consists mainly of rather pure, generally light-colored quartzite, dark gray and greenish chert, some gray to black siliceous shale, and a significant amount of greenstone. The upper member consists principally of dark thin-bedded chert interbedded with dark shale and a little greenstone. The base of the Valmy is concealed but at least 4,000 feet is present. The upper beds of the Valmy are highly contorted, but are estimated to be 3,000 or more feet thick. [Refer also to Ross (1961).]

In the shelf region the Transcontinental Arch was nearly completely emergent, or at least no Ordovician strata occur on it under a Devonian and Mississippian cover, except for the Colorado sag. This embayment probably did not extend all the way through to the western geosyncline or the Williston basin because in the eastern Uinta Mountains of Utah the Mississippian beds (possibly Devonian) rest directly on the Cambrian.

The ancestral Sweetgrass arch was broadly emergent and well-defined.

Silurian Basins

The Silurian seas were more restricted than any others in Paleozoic time. The Laketown dolomite of northern Utah and southeastern Utah has been traced widely over western Utah and is the sole representative of the Silurian thus far recognized there. In eastern and central Nevada the Roberts Mountain formation and overlying Lone Mountain dolomite correlate with the Laketown. The entire section is carbonate rock, and over half of it is dolomite (see Figs. 6.9 and 6.12).

Silurian rocks of eugeosynclinal aspect appear to be widespread in north-central Nevada, but because they resemble the Ordovician units they may not have been recognized in mapping.

On the east side of Pine Valley about 8 miles south of Carlin, unnamed black shale and tawny to buff tuffaceous shale and calcareous shale have yielded Monograptus determined by R. J. Ross, Jr., to be of Silurian age. The thickness of these beds is not known.

Black shale containing Monograptus is reported by C. W. Merriam (oral communication) from the vicinity of McClusky Pass in the northern part of the Simpson Park Mountains. C. A. Nelson (oral communication) also reports Monograptus in shale on the east side of Pine Valley near Mineral Hill. On the west side of the Tuscarora Mountains in the valley of Mary's Creek, graptolites that according to R. J. Ross, Jr., have affinities with Silurian forms were collected by Roberts in 1954. Silurian strata (R. J. Ross, Jr.), including about 4000 feet of sandstone, arkose, shale, and a little chert, from part of the overriding plate of the Roberts Mountains thrust in the northern Shoshone Range and in the Cortez Mountains.

The beds containing graptolites of Silurian age are on the whole less cherty, and contain more calcareous shale and limestone layers than the Vinini and Valmy formations. On the other hand, the Silurian beds of the western assemblage appear much less calcareous than the Silurian of the transitional

assemblage. The western rocks contain some siliceous pyroclastics, which have not been recognized in the other assemblages (Roberts *et al.*, 1958).

Silurian strata have been recognized in the northern Klamath Mountains by Wells *et al.* (1951), and rest on highly foliated schists which may be metamorphosed Ordovician and Cambrian or Precambrian in age. The Silurian beds had formerly been considered Devonian, but patches of Devonian limestone of undetermined stratigraphic relations crop out nearby. The sequence of units now recognized by Wells and co-workers is as shown in Fig. 6.13, and is compared with the assemblage of rock units in the southern Klamath Mountains. Since no Ordovician or Cambrian beds are yet known west of north-central Nevada, the possibility of correlating the Salmon and Abrams schists with the Ordovician and Cambrian is suggestive. The Copley and Chanchellula are questionably correlated with the "Silurian strata" of the northern Klamath Mountains.

Devonian Basins

The Devonian basins are in much the same pattern as the Ordovician although the strata are not so thick. The Transcontinental Arch in Utah and Arizona was more widely covered, however (Fig. 6.5).

Although Devonian strata are found nearly everywhere west of the Transcontinental Arch (Brooks and Andrichuk, 1953), they are over 1000 feet thick only in the western part of the general Rocky Mountain area. In the Roberts Range, Nevada, Merriam (1940) has described 4465 feet of Devonian beds, and at nearby Eureka he has found 4000 to 5000 feet of them. They are composed chiefly on limestones and dolomites, their fossil content indicates a rather complete section, and the broad trough in which they accumulated subsided during most of Devonian time. (See Fig. 6.9.)

Devonian rocks of the Sulphur Spring and Pinyon ranges have been recently described by Carlisle and others, who showed that northward from the Roberts Mountains the Nevada–Devils Gate sequence thickens, becomes more dolomitic, and less fossiliferous. The sequence contains vitreous quartzite units as much as 400 feet thick that grade into carbonate quartz arenites and thus resembles the Devonian section near Eureka more than the section at Lone Mountain.

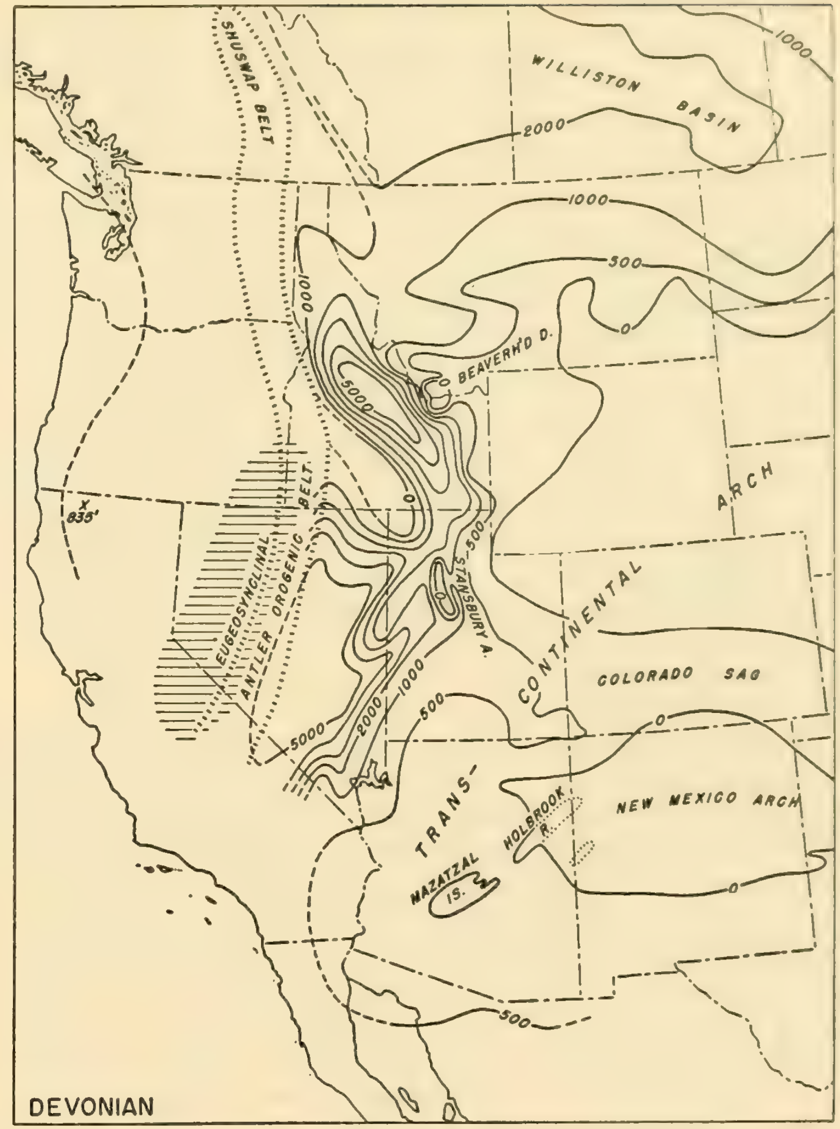


Fig. 6.5. Thickness and paleogeographic map of the Devonian. Antler orogenic belt, Stansbury anticline, and Beaverhead dome made appearance in Late Devonian. Most of sediments are Middle Devonian.

Devonian rocks of the western assemblage appear to be widespread throughout north-central Nevada, but are most abundant from the Shoshone Range eastward. These lack the basic volcanic flows and pyroclastics characteristic of Cambrian and Ordovician rocks of the western assemblage, but locally contain silicic pyroclastics, much chert and shale, and a little calcareous shale.

In Slaven Canyon in the Shoshone Range and elsewhere in the Mt. Lewis Quadrangle, there are at least 4,000 feet of strata composed dominantly of dark gray to black chert with some dark shale, a little sandstone, and very small amounts of limestone. These have yielded ostracods and conodonts of Middle Devonian age. Similar rocks on and south of Bald Mountain, in the northern Toyabe Range southwest of Cortez, are probably correlative.

Tuffaceous shale and calcareous shale on the east side of Pine Valley about 8 miles south of Carlin have also yielded conodonts of Devonian age. These rocks are associated with Silurian and Ordovician rocks in the upper plate of the Roberts Mountains thrust (Roberts *et al.*, 1958).

In the southern Klamath Mountains siliceous black shales and slates containing thin beds of sandstone and fossiliferous limestone, now largely recrystalized, make up the Kennett formation of Devonian age. It crops out in two restricted belts, and rests unconformably on the older rocks. Devonian strata are not known in the Sierra Nevada or Coast Ranges south of the Klamath Mountains in California.

Late Devonian Orogeny

Toward the end of the Devonian period, according to Nolan (1943), a geanticline began to rise in central Nevada, approximately along the transition zone of eugeosynclinal and miogeosinclinal sediments. See Fig. 6.5. The uplift divided the geosyncline into a western and an eastern trough, and the distribution of Devonian sediments is reflected in two ways, viz., by the almost complete removal of the earlier Devonian deposits along the axis of the arch, and by an eastward shift to the vicinity of Eureka, Nevada, of the zone of maximum sedimentation. The geanticline was later named the Manhattan (Eardley, 1947). Since then a large amount of significant field work has been done and the geanticline has come to be recognized as a belt of major orogeny, and has been called the Antler orogenic belt (Roberts *et al.*, 1958).

At the close of the Devonian, fundamental changes took place along the western part of the area of predominantly carbonate deposition (miogeosyncline). The carbonate rocks were folded and overridden by the Roberts Mountains thrust plate that brought clastic and volcanic rocks of equivalent age but different facies from the west or northwest. Clastic rocks eroded from the rising upland in the west marked the end of the broad geosyncline in north-central Nevada as it had existed earlier, and introduced a change to narrow straits and embayments in the orogenic belt during the remainder of the Paleozoic. The clastic rocks do not resemble the assemblages laid down in the geosyncline during early and middle Paleozoic, but overlap all of them. On the west, overlapping rocks rest with angular unconformity on rocks of the western and transitional assemblages; in the Carlin area, west of Elko, the unconformity is much less marked; and on the east, the discordance fades out and the overlapping rocks interfinger with the eastern assemblage rocks and grade eastward into the carbonate section of late Paleozoic age of eastern Nevada and western Utah. Examine Figs. 6.9, 6.14, and 6.15.

In latest Devonian or earliest Mississippian time a sharp anticline rose in the site of the Stansbury Range of west-central Utah. It was eroded down to the Cambrian before early Mississippian seas covered it (see Fig. 6.12). Coarse slide debris accumulated on its northwest flank, and sand dunes were blown northward for several miles to build a sandstone unit several hundred feet thick. The angular unconformity and the completeness of the anticline, about 30 miles long and 5 miles wide as mapped by Rigby (1958), are particularly impressive.

No Devonian strata are known in the Raft River Mountains of northwestern Utah; only Pennsylvanian strata in fault contact with Precambrian rocks have been mapped, and the Devonian relations have not been specifically deciphered (Felix, 1956). Small remnants of allochthonous Paleozoic (?) strata occur on the Precambrian rocks, and the possibility exists that this area may be a continuation of the Stansbury anticline and a belt where orogeny was more severe than to the south. The belt may join the Antler orogenic belt to the northwest. See Fig. 6.5. More details of the Antler orogenic belt will be given in the discussion of the Mississippian, Pennsylvanian, and Permian strata.

Mississippian Basins

Major miogeosynclinal deposits extend from the Big Snowy basin of Montana in a fairly narrow trough southward through eastern Idaho into Utah and then southwesterly into southern Nevada. The greatest thickness is reached in the Lemhi and Lost River ranges of Idaho (Figs. 6.6 and 6.11).

Characteristic formations of the trough are shown in Fig. 6.16. In summary of the strata of the eastern trough it may be said that they consist mostly of limestones, but that the limestones grade into a thick shale (now argillite) section in Idaho, which may savor of the eugeosyncline. Also the Manning Canyon shale of western Utah is thick (1100 feet) and marks the transition from the Mississippian to the Pennsylvanian. For references see Scholten (1957), Morris (1957), and Gilluly (1932).

The change from shelf to miogeosyncline is shown in Figs. 6.11 and 6.17. The Raft River geanticline just southwest of the Montana–Idaho boundary is well illustrated in Fig. 6.11.

Antler Orogenic and Post-Orogenic Stratigraphy

Coarse clastics in places 10,000 feet thick were spread eastward and westward from the Antler orogenic belt, and overlap the pre-existing eugeosynclinal, transitional, and miogeosynclinal assemblages. According to Roberts *et al.*, (1958):

> The lithologic character of the overlap assemblage is variable from place to place, and different names have been applied to correlative beds. In the east, the Eureka–Carlin sequence includes the Chainman shale, Diamond Peak formation, Ely limestone, Carbon Ridge, and Garden Valley formations of the Eureka area, and correlative formations in the Carlin area. In the west, the Antler sequence includes the Battle formation, Highway limestone, Antler Peak limestone, and Edna Mountain formation. Because of local variations in source areas, in conditions of deposition, and subsequent history of these rocks, it is impossible to make precise correlations of the units in the different sequences. Regional lithologic similarities indicate, however, that similar environmental conditions prevailed over broad areas. The Havallah formation of the Sonoma and East ranges was probably laid down 50–100 miles west of the orogenic belt and was thrust eastward into juxtaposition with the Antler sequence during Mesozoic orogeny. It therefore has had a somewhat different history and is

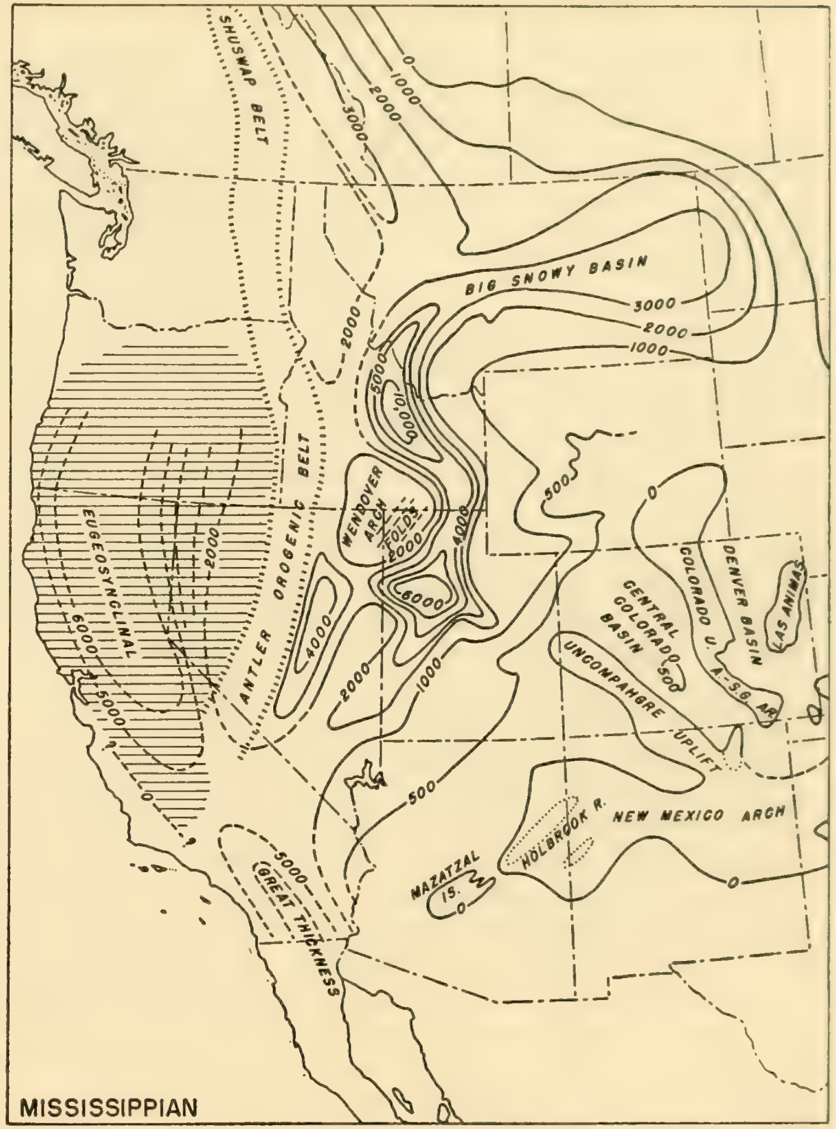


Fig. 6.6. Thickness and paleographic map of the Mississippian. A–S.G. AR. is Apishapa-Sierra Grande arch. Uncompahgre and Colorado uplifts first became emergent in latest Mississippian, and developed into major ranges in Early Pennsylvanian.

not strictly comparable with the approximately contemporaneous Eureka–Carlin and Antler sequences.

The basal sediments of the overlap assemblage differ in age throughout north-central Nevada. In the Eureka area the intertonguing Chainman shale and Diamond Peak formation of Late Mississippian age are the earliest orogenic sediments recognized. In the Carlin area the Tonka formation of Dott (1955, pp. 2222–33) and correlative units farther southeast in Pine Valley mapped by J. Fred Smith and Keith Ketner included Lower Mississippian clastic beds that overlap the upper plate of the Roberts Mountains thrust fault, indicating that the thrust reached the Carlin area during Late Devonian or Early Mississippian time.

Orogenic movements continued along the belt in Pennsylvanian and Permian time, and also throughout the Mesozoic. Examine the structure cross sections of Chapter 17, Figs. 17.3–17.6.

Walter Sadlick and F. E. Schaeffer (personal communication) recognize an angular unconformity at the base of the Chainman formation in western Utah and are calling the disturbance represented by it the Wendover phase of the Antler orogeny. They are of the opinion that this time (early Valmeyer of the early Mississippian) marks the beginning of the Antler orogeny. They recognize beveled folds covered by the Chainman, and the axes of the folds trend to the northwest.

Klamath Mountains and Sierra Nevada

The Mississippian is made up of two formations in the Klamath Mountains, the Bragdon and the Baird (Fig. 6.13). They are probably the most widespread formations in the region. The Bragdon is chiefly shale and slate, generally gray, in contrast to the black shale and slate of the older Kennett formation of Devonian age. Some sandstones are conglomeratic near the base and contain fragments of both the Kennett and Copley formations. Within the Redding quadrangle, a volcanic sequence called the Bass Mountain basalt is present. The Bragdon may exceed 6000 feet in thickness in places. The Bass Mountain volcanic sequence contains many tuff beds. Its position on Bass Mountain, according to Hinds (1939), is in the lower part of the Bragdon formation.

The Baird formation consists largely of sandstone and tuff, but the upper part has calcareous and siliceous slates. It is about 700 feet thick and apparently rests conformably on the Bragdon (Hinds, 1939).

In the northern Sierra Nevada, the metamorphic Calaveras formation of Carboniferous age is widespread. It consists chiefly of black phyllite with subordinate fine-grained quartzite, limestone, and chert. Associated and in part interbedded with the formation are green amphibolite schists of contemporaneous age. From fossils, found chiefly in the limestone, the Calaveras formation is known to be at least in part of Carboniferous age, but parts of it as mapped may be Devonian and Triassic. Because of the metamorphosed condition of the rocks in which the fossils are found, it has been difficult for paleontologists to determine to what part of the Carboniferous the faunas belong. Groups of Calaveras fossils from the Taylorsville region are more closely related to the Baird, now recognized as Mississippian, than to the McCloud limestone, now believed to be Permian.

The amphibolite schists were originally fine pyroclastics (Knopf, 1929). The bedded rocks are most abundant in the northern Sierra Nevada, but southward become increasingly metamorphosed, and progressively greater areas are occupied by granitic intrusives. In the Tehachapi Mountains and the southern Coast Ranges, pre-granitic rocks are present, but highly altered.

A thick sedimentary deposit, now schist, in southern California, has yielded Mississippian fossils (Larsen, 1948). The sequence appears to be miogeosynclinal in type and at the same time seemingly out of place in the geosynclinal setting.

Pennsylvanian Basins

Of the miogeosyncline the Oquirrh basin is the most striking feature of Pennsylvanian and Permian time. It appears to have been a sharp and small basin in which over 15,000 feet of strata accumulated. The thickest section is in the Provo part of the Wasatch Mountains of central Utah where Baker (1947) reports 26,000 feet of beds. The upper 9800 feet is of Permian age. A short distance to the southeast 20,000 feet of beds have been estimated in the Mt. Nebo district (Eardley, 1934), and in the range to the west, the Stansbury, 15,000 feet (Rigby, 1958). The basin has been contoured with a northwest trend and an abrupt northeast margin (Stokes and Heylmun, 1958). This permits the interpretation

that the Uncompahgre Range of the Ancestral Rockies (Chapter 15) extends through in subdued form to a small uplift in northwestern Utah. The sharp margin was not a fault scarp, however, because no coarse flanking debris is known as in Paradox basin. The conspicuous change from shelf to basin is illustrated in Fig. 6.17. The basin was filled, at least on the north by progressive overlap from south to north, with the oldest Pennsylvanian Morrowan sediments on the Manning Canyon shale on the south and with Atokan, Desmoinesian, and Missourian successively deposited on the shale to the north (Rigby, 1958). Limestone and sandstone are the principal lithologies in the thick succession, and cyclical sediments dominate the Desmoinesian section in the Stansbury Mountains. Quartzite and sandstone dominate over limestone in the Missourian and Virgilian section.

A deep and evidently large basin developed in Idaho in which the Wood River formation accumulated possibly 12,000 feet thick. The formation extends westward from the Lost River Range an unknown distance. The shelf deposits in southwestern Montana are represented by the Quadrant quartzite which attains a maximum thickness of 2600 feet (Scholten, 1957). The Wood River contains fusilinids of Desmoinesian, Virgilian, and Wolfcampian ages (Bostwick, 1955), and therefore was deposited simultaneously with the upper part of the Oquirrh formation.

The basal Wood River consists of several hundred feet of conglomerates, consisting of angular to well-rounded chert and quartzite pebbles. Dark arenaceous limestone beds overlie the conglomerate, and then the rest of the formation, which is the bulk of it, is a monotonous sequence of calcareous sandstones and sandy limestones. Recrystallization and replacement are common. The sandstones are mostly made up of quartz grains with 5 percent or less of feldspar, moscovite, magnetite, and zircon. The formation is characterized as miogeosynclinal by Bostwick. Although the sandstones may resemble those of the Quadrant to the east, it is difficult to see how the conglomerate could have been derived from an eastern source and transported over the region of sand deposition. It seems more logical to think of the chert and quartzite pebbles coming from the west, and thus the inference is drawn that the Antler orogenic belt extended from Nevada northward through central Idaho, and was the source of the conglomerate and, possibly, of much of the sand.

The relation of Pennsylvanian rocks to the Antler orogenic belt is diagrammed in Figs. 6.14 and 6.15.

In Nevada the Pennsylvanian rocks, like the underlying Mississippian are particularly thick east of the orogenic belt, but not quite so coarse.

Basal beds in the overlap assemblage near the orogenic belt, expecially in Mississippian and Early Pennsylvanian, are usually coarse conglomerates which grade laterally into finer conglomerates and sands, then into silt, clays, and limestone. These clastic beds may be terrestrial locally within the belt, but they are mainly marine adjacent to it. The belt may have been largely submerged at times, for widespread marine limestone units interfinger with the clastics. The lenticularity of the overlap sediments as a whole suggests deposition in several separate basins, possibly in a series of straits separated by peninsulas and islands. The presence of coarse clastics throughout much of the Pennsylvanian indicates continued orogenic activity from time to time, perhaps continuing into the Permian (Roberts *et al.*, 1958).

Volcanoes were active west of the orogenic belt as attested by the presence of volcanic materials particularly in the Pumpernickel and Havallah formations. These deposits are believed by Roberts *et al.* to have been moved as an allochthonous mass a number of miles from the vicinity of the Nevada–California border eastward to the west side of the orogenic belt, because they have no lithic counterparts nearby. The Calaveras beds in the Sierra Nevada appear to have been metamorphosed more than associated Jurassic beds (refer to Chapter 17), and since no Pennsylvanian rocks have been recognized in the Sierra Nevada or Klamath Mountains, an episode of low-grade dynamic metamorphism has been postulated in Pennsylvanian time. Accordingly on the map of Fig. 6.7 an orogenic belt is shown in the California region.

A thick quartzite formation overlies a Mississippian schist in southern California and is here placed in the Pennsylvanian although no fossils have been found in it (Larsen, 1948).

Permian Basins

The Permian was a time of extensive volcanism in the west, and various kinds of volcanic rocks were spread from the Klamath Mountains on the Pacific coast to central Nevada. The sequence is 5000 feet deep at Blairsden in the Sierra Nevada and thickens eastward to 12,000 feet in

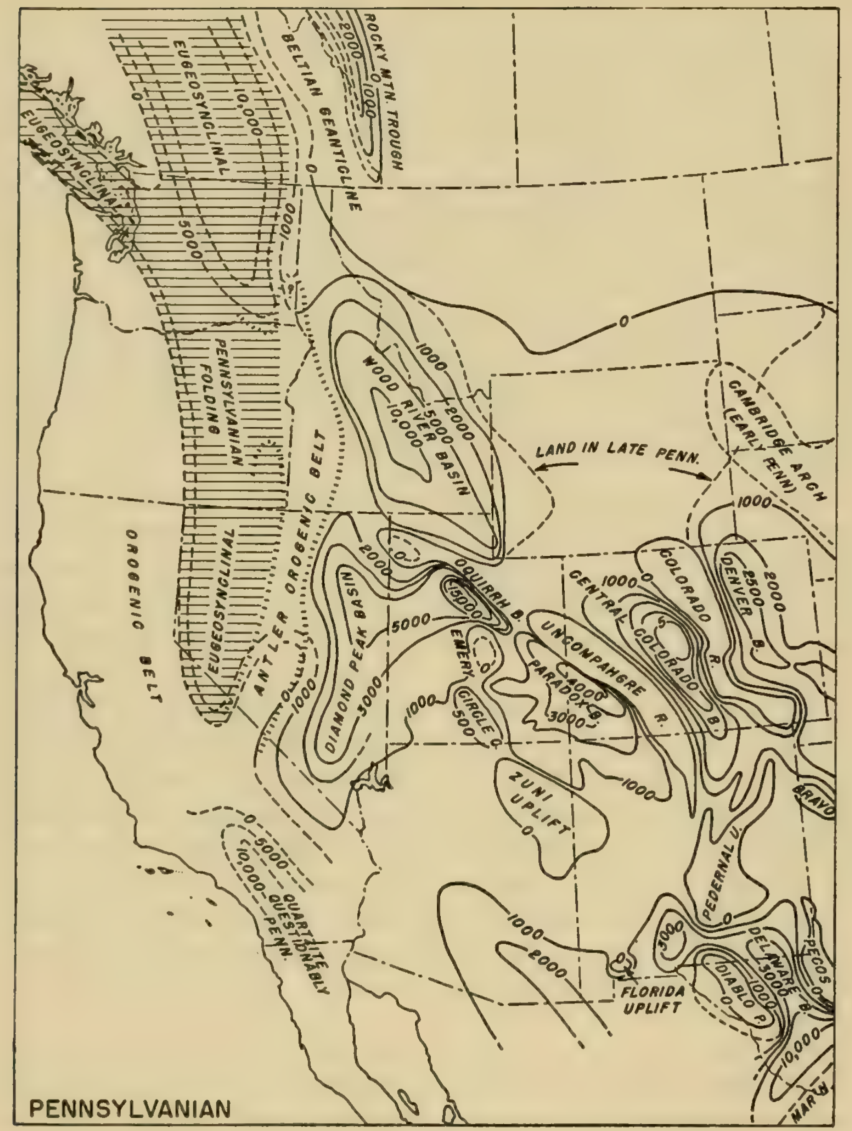


Fig. 6.7. Thickness and paleogeographic map of the Pennsylvanian.

the Humboldt Range, Nevada (Nolan, 1943). Northwestward into central Idaho, it thins to about 4000 feet. See Figs. 6.8 and 6.13.

In the Klamath Mountains the Nosoni formation occurs and is composed of basaltic agglomerates, lithic crystal tuffs, flows of andesite and of olivine basalt, dark brown, fossiliferous, shaly limestone, and dark gray to brown tuffaceous shales and slates. The maximum thickness measured by Hinds is 1200 feet. It is considered to be upper Lower Permian (Wheeler, 1933).

The Nosoni rests, probably unconformably (Hinds, 1939), on the McCloud limestone which is highly fossiliferous. It was probably a massive cherty limestone, but now owing to the Jurassic intrusions it is mostly metamorphosed in various degrees to marble. It reaches a maximum thickness of 2000 feet. Its fossils were first thought to represent a Pennsylvanian age, but a recent study by Wheeler (Hinds, 1939) shows them to be Lower Permian.

The McCloud limestone overlies the Mississippian Baird formation disconformably, so it appears that most of the Pennsylvanian was a time of emergence.

Central and Eastern Oregon. A heterogeneous group of east-west trending ranges and dissected lava plateaus known collectively as the Blue Mountains uplift or the Blue Mountains–Ochoco Mountains uplift (Waters, 1933) extends from central to eastern Oregon. The ranges are formed of Paleozoic and Mesozoic sediments and lavas and Mesozoic plutons, and the complex protrudes island-fashion through the Columbia River lava fields. The oldest beds that crop out are Lower Carboniferous limestones and calcareous sandstones (Merriam and Berthiaume, 1943). See Fig. 6.18. About 1000 feet of them are exposed, and they are called the Coffee Creek formation. No volcanic materials have been noted.

Overlying the Coffee Creek formation is the Spotted Ridge formation. The exact contact relations have not been observed, but if an unconformity does exist, it is probably not angular and does not represent much of a time break. The Spotted Ridge consists of plant-bearing sandstones and mudstones, conglomerates containing diorite, andesite, and dacite boulders, and bedded chert. It may be 1500 feet thick. The plants are believed to be Lower Pennsylvanian.

The Coffee Creek and Spotted Ridge formations are reported as intensely folded, but no mention is made of metamorphism (Merriam and Berthiaume, 1943). They lie in a tectonic belt of deformed strata in which the rocks on the west (Klamaths) and on the east (Baker area) are metamorphosed, and it is puzzling that these also are not metamorphosed.

The Spotted Ridge is overlain by the Coyote Butte formation. A slight angular unconformity separates the two. The Coyote Butte is made up almost entirely of massive limestones. Some chert pebble conglomerates are present near the base. The age is probably Lower Permian.

A prominent angular unconformity exists between the Paleozoic beds of central Oregon and the overlying Triassic conglomerates which attain a thickness of 4000 feet.

In the Baker quadrangle of eastern Oregon, Gilluly (1937b) described a formation, the Burnt River schist, which, chiefly because of greater metamorphism than that of known Carboniferous rocks nearby, he cautiously treats as older. The rock varieties are greenstone schists, quartz schist, conglomerate schist, limestone, slate, and quartzite, and make up a series at least 5000 feet thick, maybe several times as much. The various types mentioned grade into each other.

Gilluly visualizes the origin of the strata as follows:

> . . . pyroclastic material was added in amounts varying from time to time to a basin of sedimentation to which at some times sand and at others clay, with some carbonates, were being supplied. When volcanic contributions were small, the deposits were such as have yielded the quartzites and carbonaceous slates now found, but when the volcanic material increased relative to the normal terrigenous sediment the deposits were such as have yielded the intermediate rocks. At times such floods of volcanic material were contributed that practically unmixed tuff was formed.

The Burnt River schist has lithologic similarities with the Calaveras formation, but differs, it seems, in generally having greater metamorphism and an absence of chert. The Burnt River appears from published descriptions to be surprisingly similar to the Salmon schist of the Klamaths, which is probably pre-Silurian. See Figs. 6.17 and 6.18.

Above the Burnt River schist is the Elkhorn Ridge argillite about 5000 feet thick. It is probably the most widespread of the pre-Tertiary formations and is a thick series of argillite, tuff, and chert with subordinate

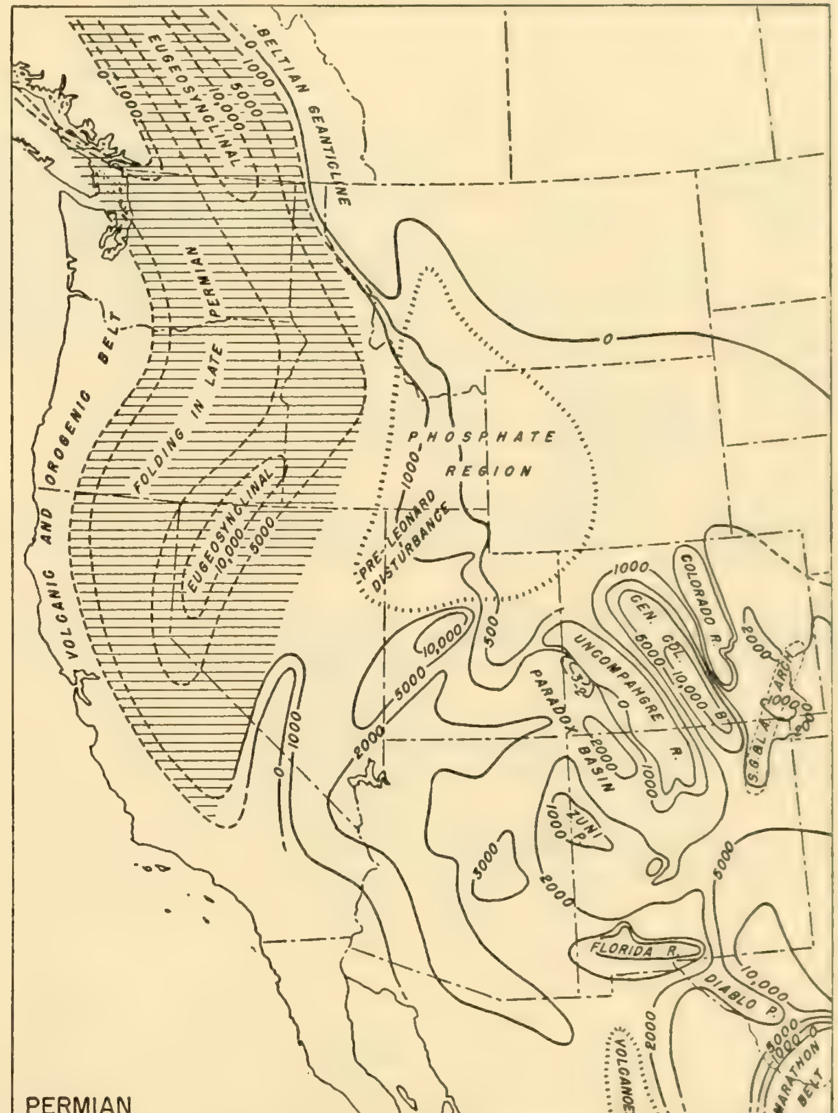


Fig. 6.8. Thickness and paleogeographic map of the Permian. S.G. and L.A. ARCH means Sierra Grande and Las Animas arch, which rose at end of Permian.

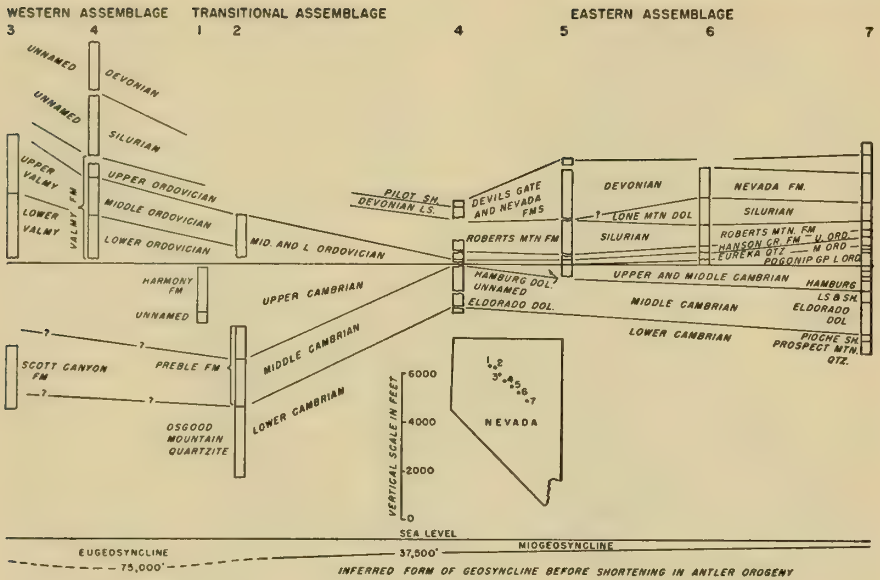


Fig. 6.9. Stratigraphic sections of pre-Late Mississippian rocks in north-central Nevada. Reproduced from Roberts *et al.*, 1958. 1, Hot Springs Range; 2, Osgood Mountains; 3, Battle Mountain; 4, northern Shoshone Range; 5, Cortez Mountains; 6. Roberts Mountains; 7, Eureka.

limestone and greenstone masses. A number of large intrusive bodies have been noted in the east-west belt of argillite; and these, together with the overlapping Cenozoic rocks, effectively prevent the recognition of contacts and the determination of stratigraphic relationships. The beds are Pennsylvanian in age, because of Fusulina fossils found in the limestones. Beds younger than Pennsylvanian may have been included in the formation as mapped (Gilluly, 1937).

The whole formation is provisionally considered marine. The tuffaceous argillite, the tuff, and the tuffaceous limestone all clearly attest notable pyroclastic contributions to the formation, and it is highly probable that cherts so numerous and thick as those in this formation may be considered evidence of igneous contribution also.

The association of limestone with volcanic materials may have no genetic significance, but a dependency is suspected because volcanism might have raised the temperature of the sea and hence decreased the solubility of the lime (Gilluly, 1937).

The Clover Creek greenstone overlies the Elkridge argillite and consists of altered volcanic flows and pyroclastic rocks, with subordinate conglomerate, limestone, chert, and argillite. It is known to extend as far eastward as the Snake River Canyon, and is therefore probably the same as the "Permian volcanics" of several areas in eastern Idaho. It is at least 4000 feet thick (Gilluly, 1937).

The effusive rocks in order of abundance are quartz keratophyre (lava-bearing albite), quartz keratophyre tuff, and meta-andesite. Fossils collected from the formation betray a Permian age.

The marine limestone and associated fossiliferous tuffs demonstrate a marine origin for part of the formation, at least. The type of albitization which most of the volcanic rocks have undergone is common in demonstrably submarine volcanic rocks, and the association here with marine limestone suggests rather strongly that the Clover Creek greenstone is in large part of submarine origin.

Northern Washington and Southern British Columbia. Where the Okanogan River crosses the international boundary, extensive areas of pre-Tertiary rocks are found. The pre-intrusives (pre-Jurassic) metamorphic rocks are called the Anarchist series; they crop out in the Okanogan Range adjacent to the Okanogan Valley on the west and extensively in the Okanogan highlands on the east. According to Krauskopf (1939) neither the top nor the bottom of the Anarchist series has been found, but at least 10,000 feet of beds exist. They can be divided rather vaguely into three divisions. The lower consists chiefly of gray to jet black phyllites with some interbedded quartzite and a little chlorite schist; the middle consists of limestone, massive quartzite, graywacke, conglomerate, some phyllite, and to the north and east of much greenstone; the upper consists for the most part of greenstone with some interbedded phyllite and quartzite. The albite in the greenstones of the upper division suggests a correlation with the keratophyres of eastern Oregon.

Regional metamorphism has converted the original sedimentary and volcanic rocks to a typical chlorite zone assemblage. Near some of the plutonic bodies higher-grade contact metamorphism has been superim-

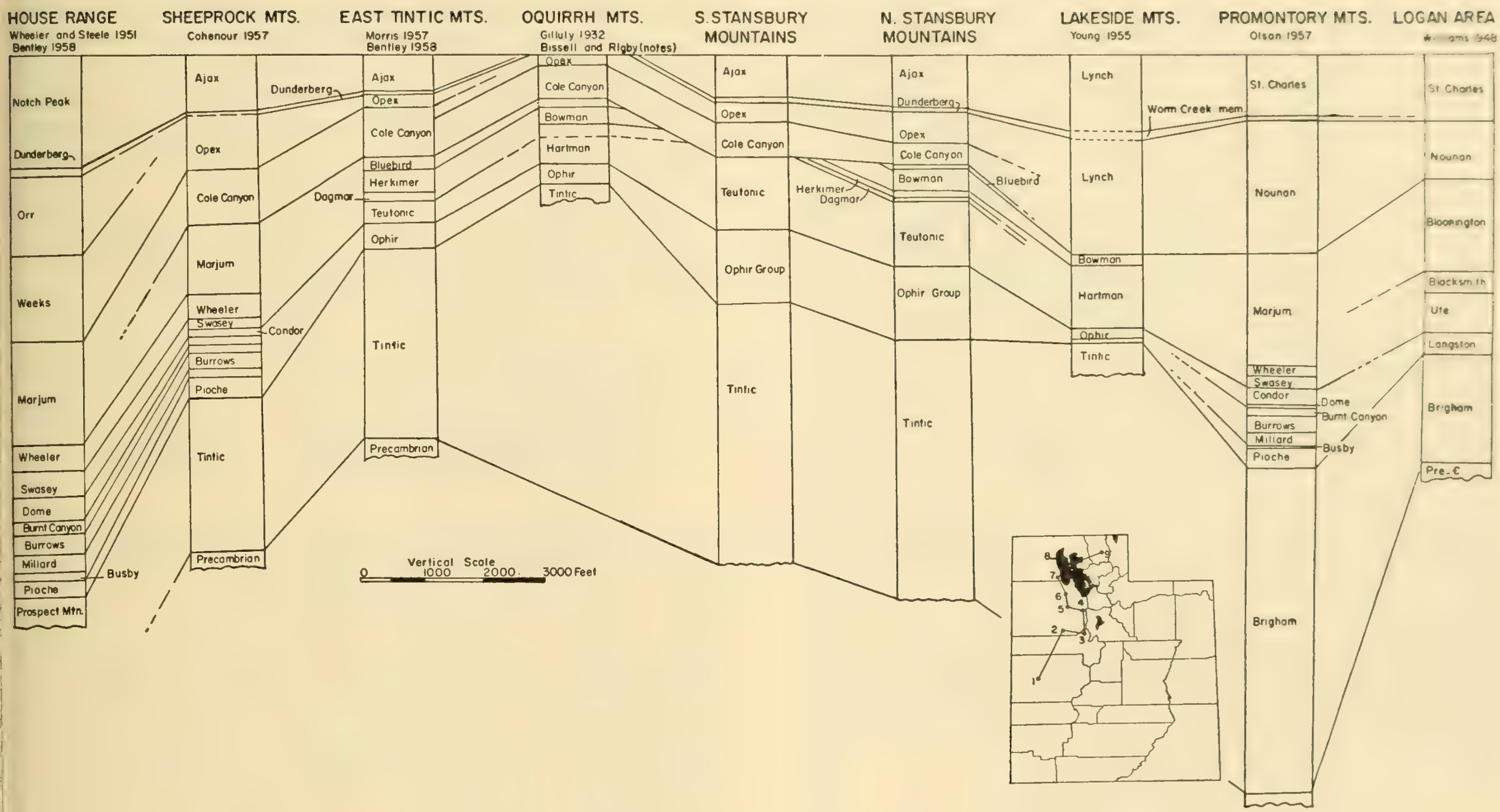


Fig. 6.10. Cambrian formations of central and northern Utah. Reproduced from Rigby, 1958.

posed, yielding biotite and amphibolite schists and diopside rocks (Krauskopf, 1939).

A few fossils establish a marine origin for part of the series at least, and a late Paleozoic age.

In Stevens County in northeastern Washington, Weaver (1920) described the Stevens series, a group of metamorphic rocks with the great thickness reportedly of 42,900 feet. It consists of quartzites, argillites, phyllites, dolomitic limestones, and schists. It is believed to be in part of Carboniferous age, but the lower parts are undoubtedly older. Bancroft (1914) had previously found fragmentary plant fossils which appeared

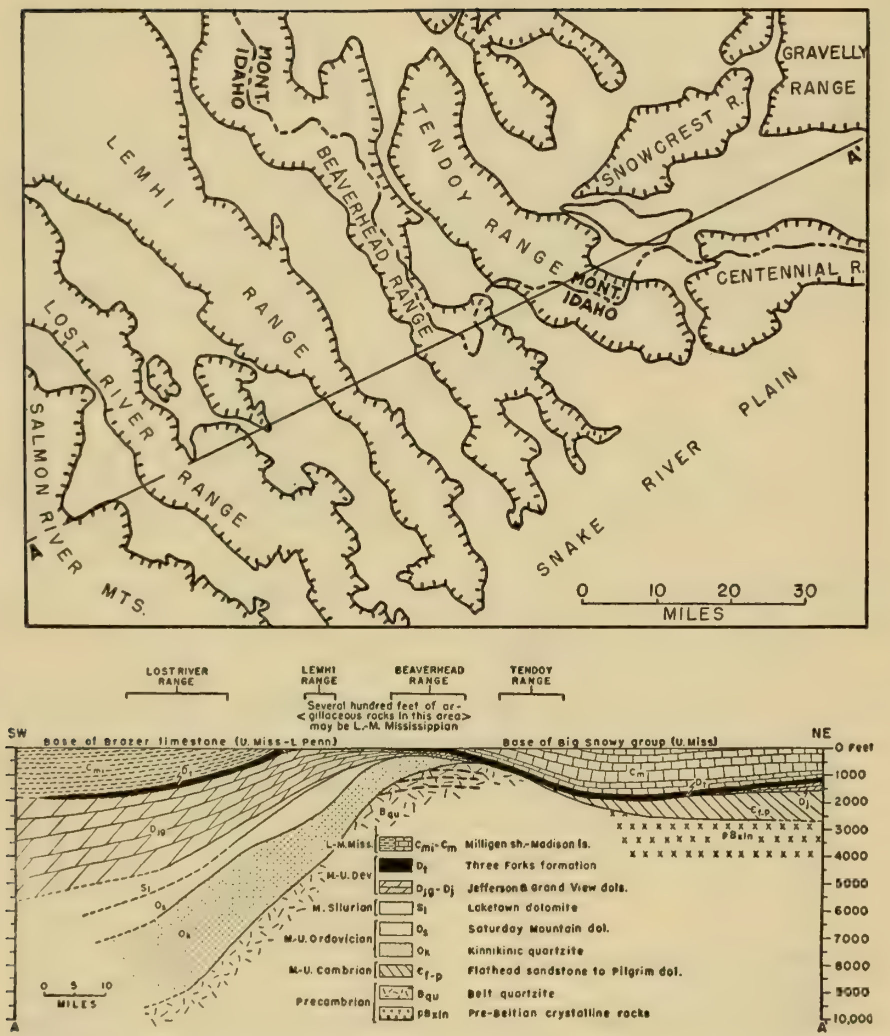


Fig. 6.11. Geosyncline, geanticline, and shelf of southwestern Montana and adjacent Idaho. After Scholten, 1957.

to be Carboniferous. The Carboniferous part of the Stevens series is probably equivalent to part of the Anarchist series on the west and to the Pend Oreille group (Daly, 1912) on the northeast along the 49th parallel. The Pend Oreille is also considered in part Carboniferous. It and equivalents rest on the immensely thick Beltian strata of Proterozoic age which form a north-south belt in northern Idaho, western Montana, and British Columbia east of Kootenay Lake.

The lower part of the Stevens series was later divided into a number of formations by Park and Cannon (1943) after Cambrian, Ordovician, and Devonian fossils had been found. Their section is as follows:

Formation	Thickness, Feet
Limestone (Devonian)	700
Ledbetter slate (Ordovician)	2500
Metaline limestone (Middle Cambrian)	3000
Maitlen phyllite (Lower or Middle Cambrian)	3000 plus
Gypsy quartzite (Lower or Middle Cambrian)	5300–8500
Monk formation (Cambrian?)	3800
Unconformity	
Leola volcanics (Precambrian)	5000 plus
Shedroof conglomerate (precambrian)	5000 plus
Unconformity	
Priest River group (Precambrian)	?

The Cambrian formations of Park and Cannon have been traced to northeastern Stevens County by Campbell (1947), where diagnostic early Middle Cambrian fossils were found.

Cache Creek Sequence of British Columbia. Upper Permian sediments are widespread and very thick over much of British Columbia (Fig. 6.8). A number of formations and groups have been collected under the general term Cache Creek sequence by White (1959). Cherts are very abundant in several forms, as well as interbedded andesite and basalt flows and related pyroclastics. Limestone units range from thin intercalated laminae to massive beds thousands of feet thick. In places the Cache Creek beds have been involved in sharp folding and metamorphism, incident to later orogenies, but where their relations to older beds have

been clearly noted, they generally rest in angular unconformity on deformed, metamorphosed, and intruded rocks.

The zone of maximum subsidence extends through the center of British Columbia with reported thickness ranging from 10,000 feet at the southern border to 24,000 feet in the northern part of the province.

The Cache Creek strata have yielded Upper Permian fossils in a number of places but lower beds in the sequence may be Carboniferous.

Shuswap Terrane and Orogeny

A large complex of metamorphosed rocks in southern British Columbia is known as the Shuswap terrane. Its location is shown on the map of Fig. 17.14. The metamorphism has long been attributed to Mesozoic batholithic processes, but now certain positive information indicates that extensive parts were metamorphosed in Pre-Cache Creek time. An authoritative summary of the nature of the Shuswap terrane by Cairnes (1939) is quoted below in which he leans toward metamorphism in Mesozoic time but recognizes that early metamorphism may have occurred.

> The rocks of this Shuswap terrane are a metamorphic complex, and their transformation is attributed to processes connected with Mesozoic batholithic intrusions, of which those of the Nelson batholith of the West Kootenay region have played a principal part. The nature of these processes is, however, not entirely clear, though certain probable conditions may be surmised from the available evidence. On the one hand it is apparent that, in part and over large areas, the Nelson batholith, together with other adjacent or comagmatic intrusives, has been emplaced to the accompaniment of much deformation in the invaded formations. On the other hand it seems equally plain that, within the broad areas occupied by much of the Shuswap terrane, the mechanics of batholithic intrusion have been of a quite different sort. There is little evidence here of those pronounced deformations with which batholithic invasion is so generally associated in mountainous regions; nor of that abrupt shouldering aside of formations flanking the irruptive mass which elsewhere characterizes the invaded strata bordering the Nelson batholith. On the contrary, batholithic invasion within the Shuswap terrane has apparently progressed under conditions of comparative stability by a process or processes of gradual soaking of the superincumbent rocks with tenuous and mobile products from the underlying magma reservoir. The nature of these products can perhaps best be judged from the occurrence of abundant bodies of pegmatitic granite throughout the Shuswap terrane; from the many associated aplitic dykes; and from the aplitic injection material that is such an important constituent of the gneissic members of the Shuswap complex. The fact, too, that large areas of massive granite contain many bodies of pegmatitic granite of precisely the same mineral composition as the granite, and show every textural gradation into these pegmatitic bodies, is further indication of the character and composition of the magmatic products effecting the transformations in the Shuswap terrane. These products are believed to have been essentially of the nature of pegmatitic and aplitic differentiates, high in volatile constituents and extremely mobile. The principal processes have seemed to involve a gradual upward seepage of this material, infiltration along bedding planes, replacement or partial replacement of intervening rock matter, and the growth, in situ, of perhaps much of the pegmatitic granite. In places the continued supply of magmatic material resulted in the complete conversion of large bodies of the original strata into massive granitoid rock which, under the conditions of transformation, became partly plastic or molten and, where subjected to local stresses, behaved much as a normal intrusive rock in its contact relations with adjoining rock masses.
>
> An important fact in the history of the Shuswap rocks, and one that has been stressed adequately by Daly, is the great depth at which their transformation has been achieved. Unquestionably the Shuswap terrane at that time was deeply buried, and unquestionably the temperatures within the zone of transformation were extremely high and long sustained. That this zone lay, in part and at times, within the zone of plastic flow is indicated in many places by numerous local sigmoid folds in which the Shuswap gneisses are involved. That temperatures within the metamorphic zone were high is indicated alone by the abundant and widespread occurrence of pegmatitic bodies everywhere within the terrane. That this condition of deep burial may, as Daly points out, afford an explanation of why the Shuswap terrane as a whole has escaped the severe deformations effecting more superficial formations (such as are now found bordering the Shuswap area), must be kept in mind in any interpretation of the origin and mode of formation of these rocks. That conditions implied by depth of burial would be most effective on the stratigraphically oldest formations is evident from the fact that for any sizable area of Shuswap rocks it is the oldest formations, or basal strata, the alteration of which has been most complete. Thus it is quite probable that within the principal area of the Shuswap terrane, as about Shuswap Lake, the formations principally effected are, as suggested also by the general structural trend of their foliation, of pre-Cambrian (Beltian?) age. In other areas, however, it is known that metamorphism has extended upward to include late Paleozoic and probably Triassic formations, but that the effects of this metamorphism have been less intense as, in general, the depth of burial has decreased.

Since 1950 evidence has been accumulating that points to the conclusion, if not the fact, that the Permian strata rest unconformably on the Shuswap and are not affected by the same orogenic and intrusive activity. Reesor (1957) summarizes recent opinion as follows:

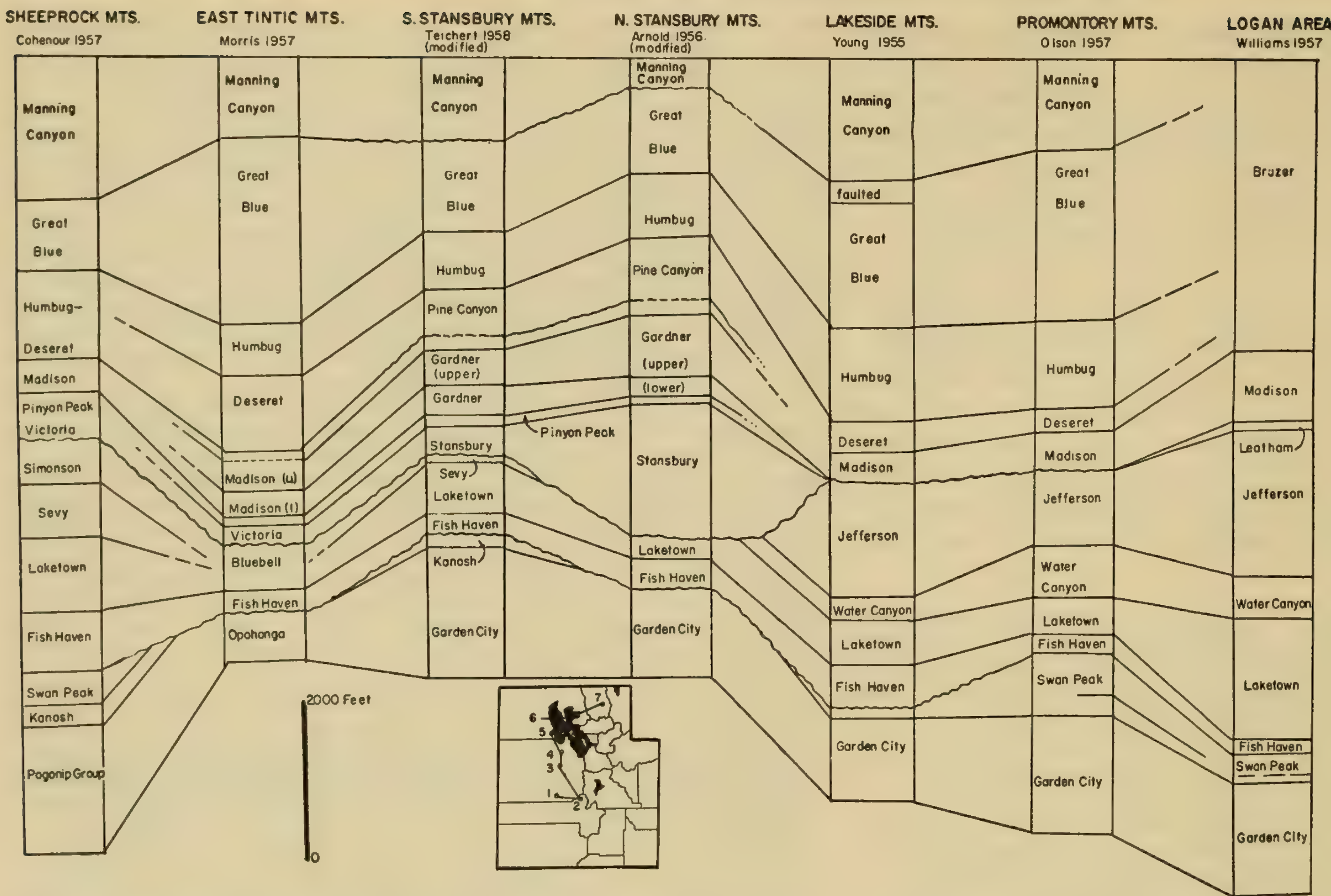


Fig. 6.12. Ordovician, Silurian, Devonian, and Mississippian formations of central and northern Utah. Reproduced from Rigby, 1958.

No reasonable doubt exists that rocks of the Cache Creek (Permian and possibly in part Carboniferous) lie with profound unconformity over rocks of the Shuswap terrain. Basal conglomerates of the Cache Creek contain boulders of metamorphosed Shuswap rocks. Thus metamorphism and deformation of the Shuswap rocks took place before the Permian.

White (1959) sites a striking example of the basal Permian unconformity in the Cariboo district. There, the Cariboo group of Early Cambrian age is closely folded into synclinoria and anticlinoria, and clastic members are regionally metamorphosed to the chlorite-muscovite grade. The Slide Mountain group of Permian age of entirely different lithology unconformably overlies the Cariboo group. It is mildly folded and not metamorphosed. Because of the clear-cut relationship here, White proposes the name, Cariboo orogeny, and includes all deformational events from Early Ordovician to Pennsylvanian in it that occurred throughout the entire Canadian Cordillera.

Attention on previous pages to Devonian, Mississippian, and Pennsyl-

vanian data in the western United States from which the maps of Figs. 6.5–6.7 were constructed lead to the suggestion that the Antler orogenic belt of Nevada extends northward through western Idaho and eastern Oregon and Washington to the Shuswap terrane of southern British Columbia. If so, we would infer that the Shuswap orogeny is of the same age as the Antler; that is, it started in Late Devonian and continued vigorously through the Mississippian and early Pennsylvanian. The Shuswap is marked by considerable metamorphism and perhaps batholithic intrusion and related processes, whereas the Antler belt is marked especially by great thrust sheets.

The term Shuswap orogeny as here used will denote tectonic events in the Shuswap terrane region that occurred during the same time approximately as those of the Antler belt, and the term Cariboo orogeny as proposed by White will be considered to have wider and longer connotation.

White (1959) summarizes information which suggests that the Shuswap belt extends into northern British Columbia and the Yukon Territory, and when so conceived the Antler and Shuswap orogenic belt is continuous from southern California to Alaska.

EUGEOSYNCLINE IN SOUTHEASTERN ALASKA, NORTHERN BRITISH COLUMBIA AND THE YUKON

Southeastern Alaska. The Paleozoic rocks in southeastern Alaska from 54° 30′ to 60° N. Lat. are of geosynclinal thickness and make up a number of formations of Ordovician, Silurian, Devonian, Mississippian, Pennsylvanian, and Permian ages (Buddington and Chapin, 1929). The stratigraphic succession is given in the table on p. 82.

One of the commonest types of rock is andesite in various forms. It occurs in at least seven formations of Permian age to Ordovician, and perhaps older. Many of the volcanic rocks are now greenstone schist. Pillow lava is abundant in the Lower and Middle Ordovician, Silurian, Middle and Upper Devonian, Lower Permian, and Upper Triassic.

The other predominant rock types are sheared graywacke, slate, and phyllite. The vast amount of greenish graywacke with associated slate is the most striking feature of the stratigraphic sequence of southeastern Alaska. Graywacke is found in every system of the Paleozoic and Mesozoic, and in many places it is difficult or impossible to tell one graywacke unit from another.

	SOUTHERN KLAMATH MOUNTAINS		NORTHERN KLAMATH MTS.
	REDDING QUAD. (HINDS *ET AL.*)	WEAVERVILLE QUAD. (HINDS *ET AL.*)	WELLS *ET AL.*
U. TRIASSIC			APPLEGATE GROUP, METAVOLCANICS AND METASEDIMENTS. FORMERLY CALLED DEVONIAN OR CARBONIFEROUS
PERMIAN	NOSONI VOLCANICS McCLOUD LIMESTONE		
MISSISSIPPIAN	BAIRD FM. BRAGDON FM. WITH BASS MOUNTAIN BASALT	BRAGDON FM.	
DEVONIAN	KENNETT FM.		DEVONIAN LIMESTONE PATCHES
SILURIAN	COPLEY META-ANDESITE (POSSIBLY SILURIAN)	COPLEY META-ANDESITE CHANCHELLULA FM. (POSSIBLY SILURIAN)	SILURIAN STRATA (NOT NAMED)
PRE-SILURIAN		SALMON SCHIST ABRAMS SCHIST	HIGHLY FOLIATED SCHIST

Fig. 6.13. Correlation of Paleozoic formations in Klamath Mountains.

Limestone forms a very considerable part of each Paleozoic formation except the Ordovician. The thickest unit is in the Upper Silurian and is a very high calcite variety. Some limestone carries considerable chert.

Beds of cobble and boulder conglomerate form conspicuous and thick members of the Silurian and Devonian formations. A peculiar but common form is composed of andesite and limestone pebbles and cobbles in a tuffaceous matrix. The same lithology is found in the Middle Devonian. Coarse conglomerate beds occur at the base of the Devonian.

Another characteristic lithology in the Paleozoic systems in southeastern Alaska is coarse, waterworn intraformational limestone conglomerate. Beds occur in the Silurian, Devonian, Permian, and Triassic formations, and in all of them the cobbles of limestone carry the same fauna as the formation in which the conglomerate occurs. Buddington believes the intraformational conglomerates originated from crustal movements accompanying the volcanic activity during these periods.

Black slate and argillite are widely distributed, and thin-layered black chert several hundred feet thick occurs in the Ordovician and Missis-

Series	Character	Thickness, Feet
	Unconformity	
Jurassic	Andesitic rocks, including breccia, with limestone matrix and lava flows (in part with pillow structure), locally interbedded with slate and other sediments	1400 plus
	Unconformity	
Upper Triassic	Conglomerate, sandstone, and limestone; in the Ketchikan district includes considerable black slate in upper part	1600 plus or minus
	Unconformity	
Permian	Thick-bedded limestone; with common to abundant intercalated layers of white chert	1000
	Conglomerate, limestone, sandstone, andesitic and basaltic lava, tuff, and locally rhyolitic volcanic rocks	3000 plus or minus
	Unconformity	
Pennsylvanian (?)	White massive limestone	100 plus
Mississippian	Interbedded coarsely crystalline limestone and black chert, overlain by interlayered dense gray quartzite and cherty limestone; sparse conglomerate	1000
Upper Devonian	Basalt, andesite (in part pillow lava), tuff, limestone, sandstone, slate, and conglomerate	1000
	Unconformity (?)	
	Limestone	600 plus
Middle Devonian	Andesitic green to gray tuff (locally cherty) and graywacke, with locally fine, conglomeratic layers, intercalated limestone, and a minor amount of andesitic lava and breccia	2400 plus
	Andesitic lava (in part pillow lava), breccia, tuff, conglomerate and locally rhyolitic lava	2000
	Interbedded limestone, slate, chert, andesitic lava, breccia, tuff, and locally conglomerate	
	Conglomerate and graywacke-like sandstone, with locally interbedded limestone	2000
	Unconformity	
	Green-gray graywacke with sparse conglomerate beds. Interbedded red, green-gray, and gray graywacke, like sandstone with small amount of shale	5000 plus
Silurian	Green-gray shale with intercalated red beds and thin-layered fine-grained gray sandstone, shale, and dense limestone	500 plus
	Predominantly thick-bedded dense limestone; intercalated with thick beds of coarse conglomerate, thin-layered limestone, nodular and shaly argillaceous limestone and sandstone Ls, 3000; Congl. 1500	4500 plus or minus
	Andesite (in part pillow lava) and andesite porphyry lava; conglomerate; with some associated graywacke, tuff, breccia, and limestone	3000 plus or minus
	Unconformity (?)	
	Indurated graywacke with associated black slate and sparse conglomerate and limy sediments	?
	Unconformity (?)	
Middle Ordovician	Indurated graywacke with associated black slate and sparse conglomerate and limy beds; locally andesitic pillow-lava and volcanic rocks	?
Lower Ordovician	Thin-layered black chert with black graptolitic slate partings, graywacke, and locally andesitic volcanic rocks	?
Probably pre-Ordovician to Devonian Wales group (metamorphic rocks)	Greenstone schist with intercalated or interbedded limestone	?
	Limestone	?
	Schist with beds of limestone and slate	?
	Schist	?

sippian formations. Thick-bedded chert and cherty tuff occur in the Middle Devonian, and white chert is common in the Upper Permian.

Schists and gneisses are also common, and are the result principally of permeating hot solutions attendant upon the emplacement and solidification of the vast volume of magma in addition to orogenic stresses (Buddington and Chapin, 1929).

Northern British Columbia and the Yukon. The *Geologic Map of Canada* summarizes what is known of the distribution of Paleozoic rocks in northern British Columbia and the Yukon. Great areas are still marked "Paleozoic, mainly sedimentary rocks," but other large areas are labeled "Carboniferous and Permian sedimentary rocks." *Geology and Economic Minerals of Canada, 1947,* summarizes the distribution as follows:

> During the Carboniferous and Permian periods apparently nearly the whole of the Western Cordilleran region (west of the Rocky Mountain trench) lay beneath the sea, and great thicknesses of sedimentary and volcanic material

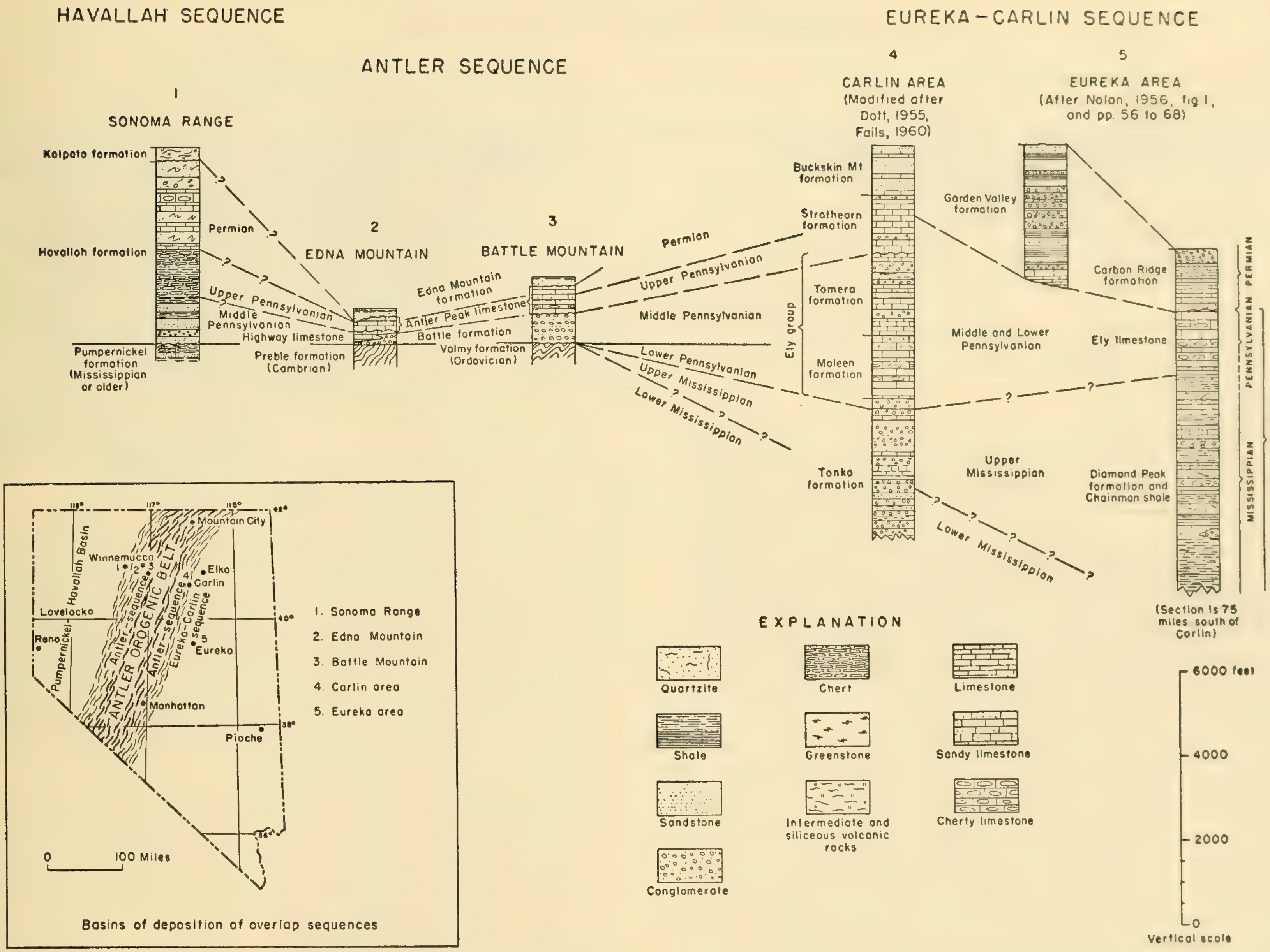


Fig. 6.14. Detail of Mississippian, Pennsylvanian, and Permian formations involved in Antler orogeny of north-central Nevada. Reproduced from Roberts *et al.*, 1958.

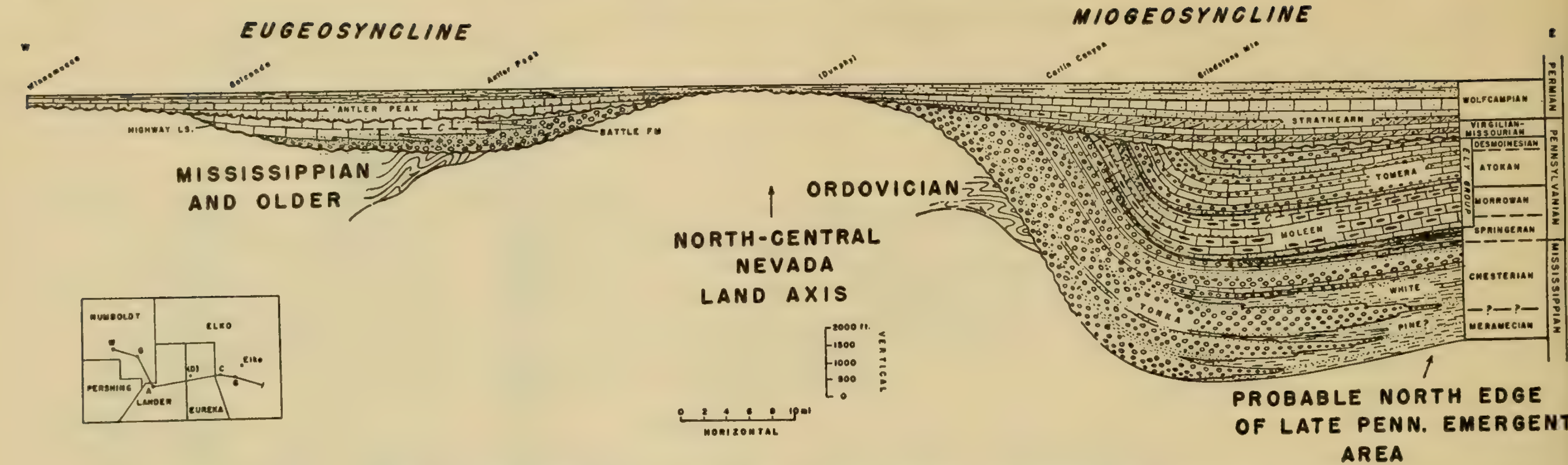


Fig. 6.15. Antler orogenic belt of central Nevada showing Mississippian, Pennsylvanian, and Permian strata restored to early Wolfcampian time. Section extends from Winnemucca to Elko. Reproduced from Dott, 1955.

accumulated. Wide areas of almost unexplored country in eastern Yukon are presumed to be underlain chiefly by Paleozoic strata, but may also contain rocks of Mesozoic and Precambrian age. In northern British Columbia, where exposed strata are thought to represent much of Paleozoic time, no important disturbance has been recognized. In a number of localities sedimentation and volcanism probably proceeded more or less continuously from late Paleozoic into early Mesozoic time, but in places an interval of uplift and erosion without marked tilting or folding may have intervened.

A report on the Cassiar Mountains, Finley River district between latitudes 56 and 58, and longitudes of 124 and 126, by Dolmage (1928) describes a series of metamorphosed rocks of Carboniferous age. They are "green ash rocks pressed and altered into schists, interbedded with layers of graywacke, felsite, halle-flinta, serpentine, and argillite." Along Takla and Stuart lakes and vicinity the series is made up of limestones, argillites, cherty quartzities, green schists, slates, volcanic flows, tuffs and breccias, and narrow bands of dolomite. Fusulina and other Carboniferous fossils have been found in some of these beds.

Underlying the Carboniferous series, great belts of schist and quartzite occur. Quartz mica schist constitutes about three-fourths of the whole. In many places, the schist grades into quartzite, both of which were derived undoubtedly from siliceous sediments (Dolmage, 1928). Such rocks as these are widespread and have been correlated with the Shuswap terrane of southern British Columbia, which now as previously explained, is believed to be made up of rocks of several Paleozoic periods as well as Precambrian. Also some coarse quartzites, quartz pebble conglomerates, and limestones have been likened to the Cambrian strata of the southern Canadian Rockies, previously described.

The areas of such rocks are shown on the map of Fig. 33.12. A great medial area of Proterozoic (Beltian?) rocks separates the western areas of Carboniferous rocks from the eastern Paleozoic rocks, but whether or not this was a highland in Paleozoic time is unknown.

SUMMARY OF OROGENIC HISTORY

The maps, Figs. 6.1 to 6.8, are fairly expressive of our present knowledge and postulates of the evolution of the western margin of the con-

tinent during Paleozoic time. That the western margin has a belt of major orogeny with associated intrusive and extrusive igneous activity and metamorphism needs no longer to be defended. At the time of writing of the first edition of this book the profession was just accepting the view and abandoning the older one of a small continental borderland, now partly submerged beneath the Pacific Ocean.

It may be stated that we have no information on conditions in Cambrian time west of northwestern Nevada. Cambrian strata are recognized farther south in California in the Death Valley region, but these lie on the projection of the eastern miogeosynclinal assemblage. Ordovician rocks, like the Cambrian, are not known for sure west of northwestern Nevada. In southeastern Alaska, however, they have been identified very close to the Pacific margin of the continent, and are part of an extensive eugeosynclinal assemblage. Silurian rocks have now been recognized near the Pacific in the Klamath Mountains, but the paleogeography of the entire region from northwestern Nevada to the Pacific is practically unknown. The presence of Silurian strata in the Klamath Mountains and sequences under them which might be Ordovician and Cambrian lead to the conclusion that the western margin of the continent as early as Cambrian time was about where it now is; and that the continent has not grown appreciably since.

We must also postulate several phases of major orogeny together with the accumulation of eugeosynclinal sequences in adjacent and associated basins or troughs in early Paleozoic times along the western margin of the continent. The transitional zones of the eugeosynclinal and miogeosynclinal assemblages are now fairly well positioned, and the basins of the miogeosyncline are beginning to take on specific shape and distribution in light of our present knowledge. Geanticlines, the Beltian and Raft River, are postulated, and the Tooele arch seems clear. These add complexity to what was previously considered a simple broad basin.

A major and unsolved problem is the relation of the southwesterly trending Paleozoic tectonic elements in southern Nevada, Arizona, and California to the continental margin—they are distinctly discordant rather than approximately concordant or unilateral. The problem has been discussed in Chapter 5.

BAYHORSE QUAD. IDAHO	LOST RIVER RANGE IDAHO	TENDOY RANGE S. W. MONT.	NORTHEASTERN UTAH	OQUIRRH MTS., CENTRAL UTAH
MILLIGEN ARGILLITE	"BRAZER"*	AMSDEN FM.* BIG SNOWY GR.	MANNING CAN. SH.* GREAT BLUE LS. HUMBUG FM. DESERET LS.	MANNING CAN. SH.* GREAT BLUE LS. HUMBUG FM. DESERET LS.
	MILLIGEN ARGILLITE	MADISON: MISSION CAN. LS. LODGEPOLE LS.	MADISON LS.	MADISON LS.
		SAPPINGTON FM.+	LEATHAM	

* PARTLY L. PENNSYLVANIAN
+ OCCURS EASTWARD AS PART OF SHELF SEQUENCE

Fig 6.16. Correlation of Mississippian formations of southwestern Montana, eastern Idaho and northern Utah.

A major orogenic belt began to develop in central Nevada in late Devonian time, and through several phases of folding and thrusting continued development through the rest of the Paleozoic. The belt is projected northward through eastern Oregon and Washington into southern British Columbia in Mississippian and Pennsylvanian time to the Shuswap orogenic belt in British Columbia. Another orogenic belt lay to the west in Pennsylvanian and Permian time, and it seems to have been separated from the central Nevada belt by a basin of sedimentation. The entire region including both belts and the intervening basin become involved in orogeny, volcanism, and intrusive activity thereafter, starting in Permian time.

Shifting basins and the appearance of uplifts of several kinds add complexity to the miogeosyncline and its relation to the shelf in the late Paleozoic.

The Canadian cordillera is not as wide as that of the western United States, and perhaps its development is more regular. From what is known it appears that a geanticline of Beltian strata developed early in Paleozoic time and separated a western eugeocynclinal trough of sedimentation from an eastern miogeosynclinal trough. The eugeosynclinal region was subjected to repeated orogeny, metamorphism, and igneous activity. In this connection it is pertinent to review Buddington's observations in southeastern Alaska.

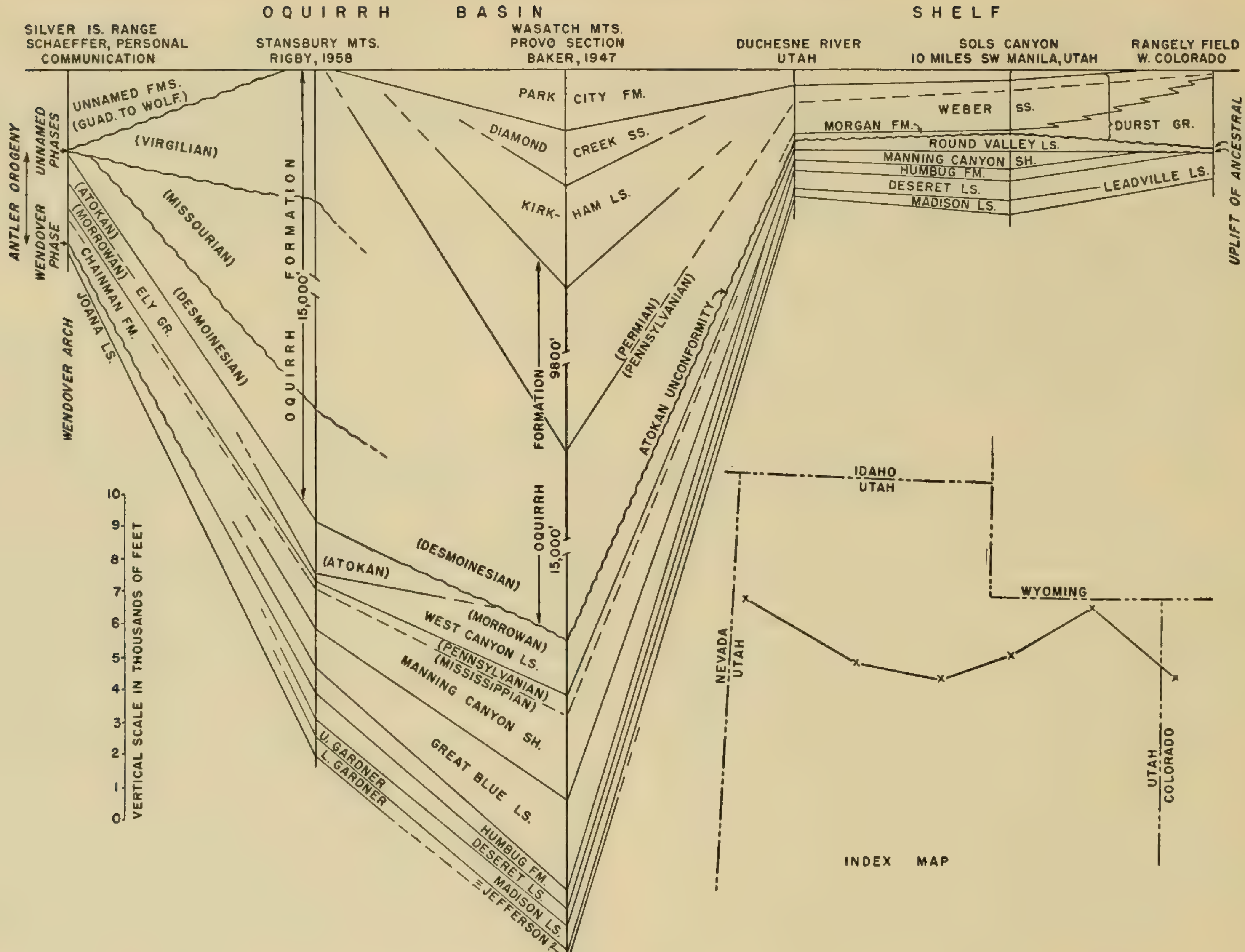


Fig. 6.17. Mississippian, Pennsylvanian, and Permian formations in Utah showing change from the shelf assemblage to the miogeosyncline assemblage. Sections of shelf were furnished by Walter Sadlick, who also assisted in the general correlations.

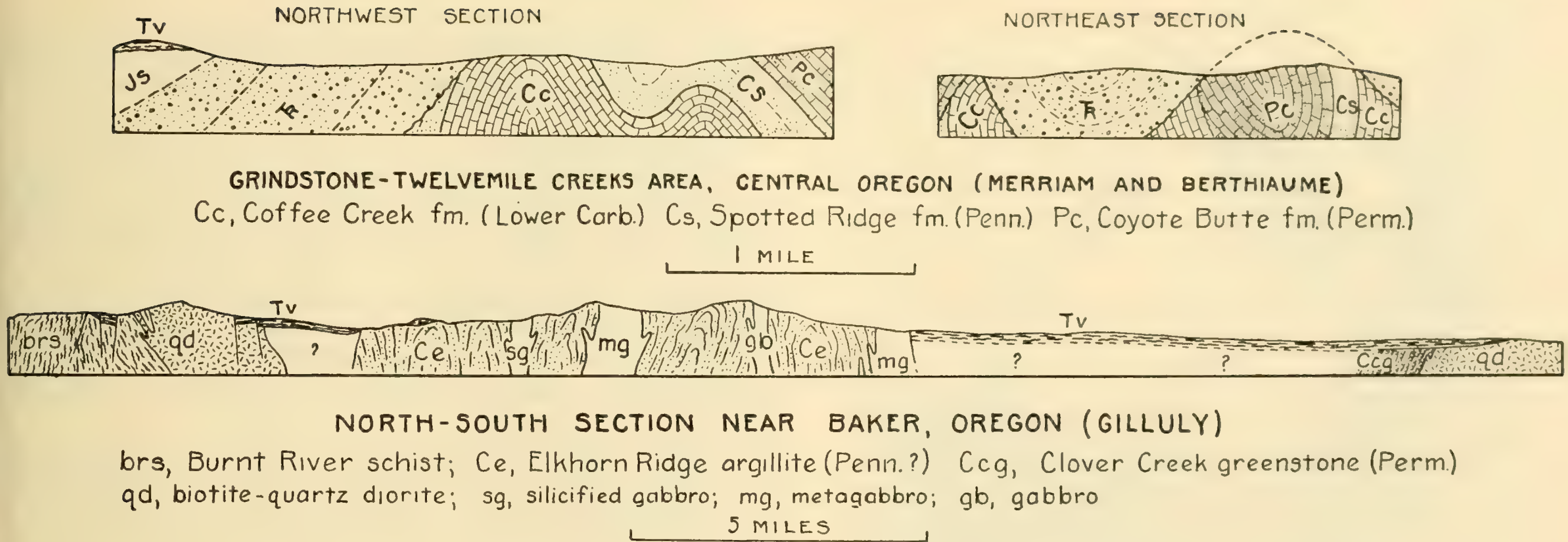


Fig. 6.18. Cross sections in central and eastern Oregon.

The Silurian graywackes in general of southeastern Alaska are composed of particles of rock similar to the kinds that form the pebbles and cobbles in the conglomerates with which they are interbedded, and in addition, of a considerable percentage of plagioclase, potassic feldspar, and quartz grains. The conglomerates are largely made of andesite pebbles and boulders, but slate, diorite, rhyolite, and limestone pebbles are abundant, if not dominant, in some conglomerates. One specimen of graywacke of Devonian or Silurian age, for example, consisted of particles of andesite, felsite, plagioclase, granophyre, quartz, spherulitic rhyolite, and orthoclase, with a chloritic and slightly calcareous groundmass.

The association of the graywackes and conglomerates that Buddington describes is very revealing of their origin. The conglomerates in themselves are indicative of a volcanic archipelago and deserve further mention. The following is a réumé of the Silurian conglomerates according to Buddington. Varieties of conglomerates are as follows:

1. A conglomerate composed almost wholly of well-rounded andesite or andesite porphyry cobbles and boulders; the matrix may be calcareous, and lenses of limestones are intercalated but limestone cobbles are sparse.
2. A conglomerate composed almost wholly of limestone cobbles or boulders in a limestone or andesitic tufflike matrix; this type is rare, but beds 100 feet thick have been noted.
3. Peculiar conglomerates intermediate between 1 and 2, consisting of pebbles and cobbles of andesite and limestone in a greenish tufflike matrix.
4. A homogeneous-appearing rock composed of fragments of andesite in a matrix of the same material; the structure is that of a conglomerate or water-worn breccia.

The limestone fragments are usually of a dense-textured limestone typical of the Silurian, and many carry fossils of Silurian age. The fossils are the same as from the overlying limestone. It is, therefore, believed that the limestone conglomerates are intraformational and that the limestone fragments are of practically the same age as the volcanic fragments. Vertical movements of the sea bottom, perhaps local, must have accom-

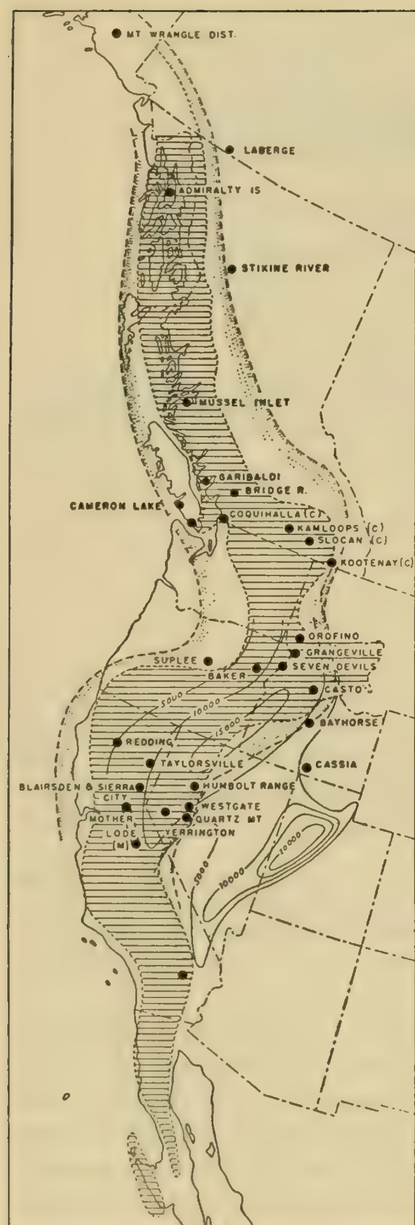


Fig. 6.19. Map showing coincidence of Permian volcanic trough (stippled margins) and zone of Sierran intrustives (lines). Dots indicate location of Carboniferous, Permian, and Triassic areas referred to the text. Pennsylvanian and Permian basins combined isopached. Zone of Sierran intrusives includes nearly all satellites and palingenetic areas.

panied the volcanism and resulted in contemporaneus erosion and submarine slumping of slightly compacted fine lime mud. A part of the volcanic material, at least, must have been erupted from central volcanoes, which were built up above the surface of the ocean and were thus subjected to erosion.

Although recognizing unsolved elements in the problem of the origin of the graywackes, conglomerates, and limy argillaceous beds, Buddington visualizes a sedimentary environment as follows: the great lens-shaped beds of conglomerate may be local deposits made by torrential streams, and the graywacke may be in part the more finely comminuted peripheral marine equivalent. The calcareous shale and argillaceous limy beds which are locally intercalcated with the clean, thick-bedded limestone may be in part the more distant offshore equivalent of the conglomerate and graywacke.

The limestone is in part dense white on fresh surfaces, and massive with only rare, if any, evidence of stratification. Beds as thick as 2000 feet have been observed. In part it is interbedded with thin-layered limestone, nodular and shaly limestone, calcareous shaly argillite, dense platy siliceous layers, green-gray shale, and sparse buff-weathering sandstone. The massive limestone seems to be due to rapid deposition, and where clean the site of accumulation was sufficiently distant from land so not to have received any clastic material. Volcanic activity has been thought of as contributing to the deposition of the limestone, through the activity of magmatic waters or meteoric waters draining from a volcanic terrane or by the warming of the marine water, but the chemistry and oceanography of the problem have not been worked out.

Schofield (1941) discussed the problem of granitoid pebbles and cobbles in the conglomerates of several periods, especially the Triassic. Buddington refers to them also. In one locality, the Britannia map area of British Columbia, an arkose is described as composed of irregular grains of quartz, plagioclase, orthoclase, and sericite schist. The lack of rounding of the grains, the freshness of the plagioclase, and the considerable thickness of the unstratified beds, prove that the material accumulated rapidly and was transported only a short distance from a source of granitoid plutonic rocks. Buddington failed to trace the granitoid clastics

to their source, despite the fact that their size and abundance indicated to him a nearby local origin. It seems necessary, he believes, to assume that granitoid intrusions existed in a land that formerly stood to the west where only the Pacific Ocean now lies.

Krynine (1941) has studied the tectonic significance of arkoses, and concludes that they are deposited when a granitoid terrane has just been uplifted and is being vigorously dissected. They are related to the deformed geosyncline into which granitoid rocks have been intruded. The plutons have become exposed by erosion of the mountains created by the orogeny, and then uplifted in a further stage of deformation, and vigorously eroded.

Granite plutons are seldom exposed in arcs of small volcanic islands. We must look to the larger islands of an archipelago for the source of granitoid conglomerates and arkoses. The geologic map of the Japanese Archipelago, Fig. 6.20, shows extensive areas of granitic intrusions and Precambrian gneisses which could furnish the necessary material. The major archipelago like the Japanese has had a long orogenic history and is composed not only of rocks that will make graywackes but also arkoses. Such a one seems to have been the sourceland of the sediments of the western part of the Cordilleran geosyncline.

Great beds of chert are present in the sediments of the volcanic archipelago. Extensive beds of chert and cherty limestone are present in the miogeosyncline as well as in the inland basins and shelfs of the mainland, and so the factors governing the precipitation of the silica are probably several. Its transportation in solution in marine currents may result in precipitation a great distance from its source. I find it easy to believe that a large part of the silica originated in the volcanic activity of the archipelago, that some of it was carried by currents across the seas between the archipelago and the mainland free from the area of deposi-

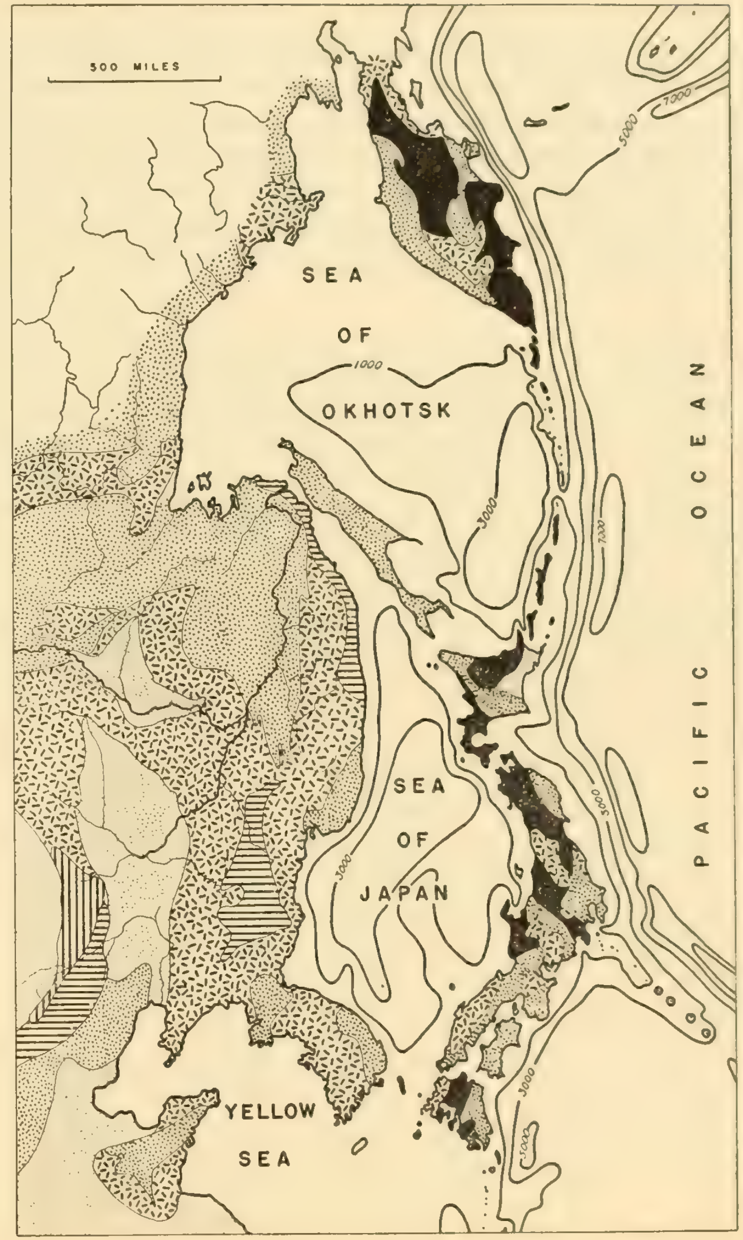


Fig. 6.20. Generalized geologic map of the Japanese Archipelago and the easten part of Asia. Isobaths are in meters. Coarsely stippled areas are those chiefly of sedimentary rocks but with large areas of Archean gneiss and schist and some smaller areas of intrusive and extrusive rock. Finely stippled areas denote alluvium. Hachured areas are those of plutonic rocks, chiefly granite and granadiorite, but with considerable areas of Archean gneiss and schist and some sedimentary rocks. Solid black areas are andesite. Horizontally ruled is basalt and verticlly ruled is trachyte.

tion of volcanic material, and that it was precipitated copiously in the shallow seas of the eastern trough and mainland shelf where, from place to place and time to time, clay, lime mud, or sand were accumulating.

Perhaps the tectonic conditions of the eugeosyncline of the western margin of North America can be visualized better if reference is made to the Japanese Archipelago (see Fig. 6.20). It is believed that there a fairly good example and close parallel of conditions exists now as existed in times past along the west coast of North America.

In the first place the scale and shape of the arcuate features are the same. In the second place, the geology of the Japanese Archipelago is somewhat the same as that postulated for the sourceland of the sediments of the Pacific trough of the Paleozoic Cordilleran geosyncline. The most abundant rocks mapped in the Japanese Archipelago are as follows: andesite, granite, syenite, schistose granite, gneiss, schist, slate, chert, sandstone, limestone, diorite, pyroxenite, amphibolite, gabbro, and trachyte, in approximate descending order of abundance.

7.

APPALACHIAN MOUNTAINS

MAJOR STRUCTURAL DIVISIONS

The index map of Fig. 7.1 shows the structural divisions of the eastern margin of the continent from New York to Alabama. The interior stable region of the continent is represented by the Appalachian plateaus province, where the strata are nearly horizontal and dissected by an elaborate arborescent drainage system. In southern New York, central Pennsylvania, and northern West Virginia, the strata are cast into very gentle folds which are the site of extensive gas and oil fields. The folding is so gentle that the drainage is little affected, and the arborescent plateau type exists, scarcely distinguishable from the region farther west.

The folded and thrust-faulted province represents the Appalachian Mountains proper. It is the well-known region of flat-topped, parallel, or subparallel ridges and valleys that are carved out of anticlines, synclines, and thrust sheets. The drainage pattern is rectangular (trellis), and stands conspicuously apart from the arborescent pattern of the Appalachian plateaus. The strata are of Paleozoic age in both provinces but thicken from the shelf along the western margin of the plateaus to the geosyncline in the eastern part of the plateaus and in the folded and thrust-faulted belt. See Fig. 7.2.

The Blue Ridge province is made up of Cambrian and Late and probably Early Precambrian metamorphic and igneous rocks, which are older than those of the Appalachians to the west, and are more or less metamorphosed. It is widest in the south, and highest in the Great Smoky Mountains of Tennessee and North Carolina (Fig. 7.1). It dies out in southern Pennsylvania only to take up again in eastern Pennsylvania, New Jersey, and New York. The Blue Ridge province is generally one of conspicuous relief east of the Great Valley of the folded Appalachians and west of the crystalline Piedmont. The Piedmont province is broad and generally of low relief. Its rocks are not well exposed and, as yet, thoroughly known in only a few places. They are chiefly metamorphosed Precambrian and Paleozoic sediments and volcanics, and Paleozoic plutons, a number of which are of batholithic proportions.

Several long, narrow basins of Triassic sediments rest unconformably on the older rocks of the Piedmont, and in one place on the Blue Ridge belt. They are down-faulted troughs, all apparently part of a major fault or rift zone. The Triassic sediments are mostly red standstones and shales, and are cut by numerous large dikes and sills of diabase, also of Triassic age.

The Atlantic Coastal Plain is a continuation of the Gulf Coastal Plain, and is made up of Cretaceous and Tertiary sediments that rest unconformably on the older rocks of all the structural systems of the Appalachian Mountains. They overlap the Triassic deposits slightly in New Jersey. They dip gently seaward and probably extend out under water

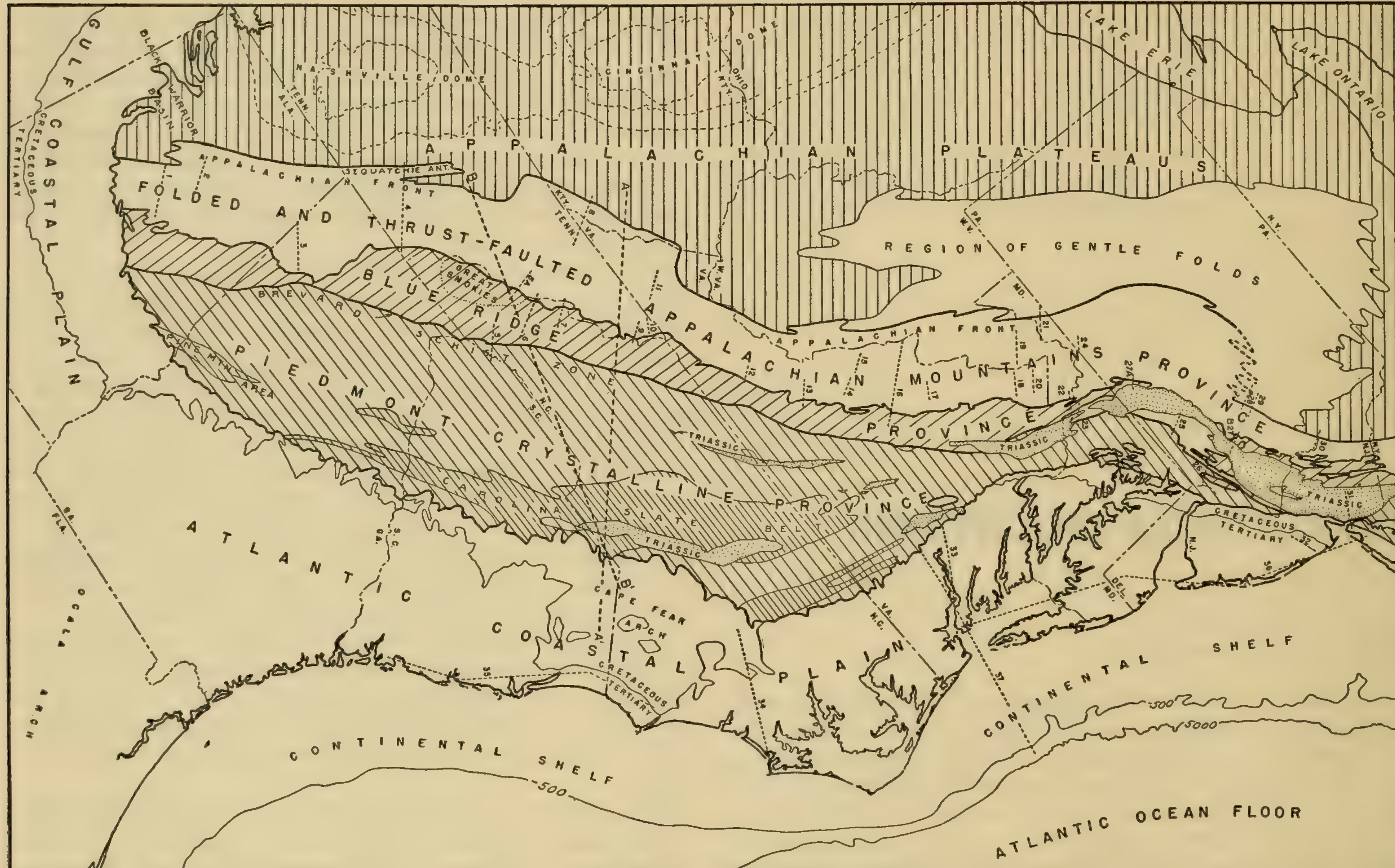


Fig. 7.1. Index map of the structural systems of the eastern margin of the continent.

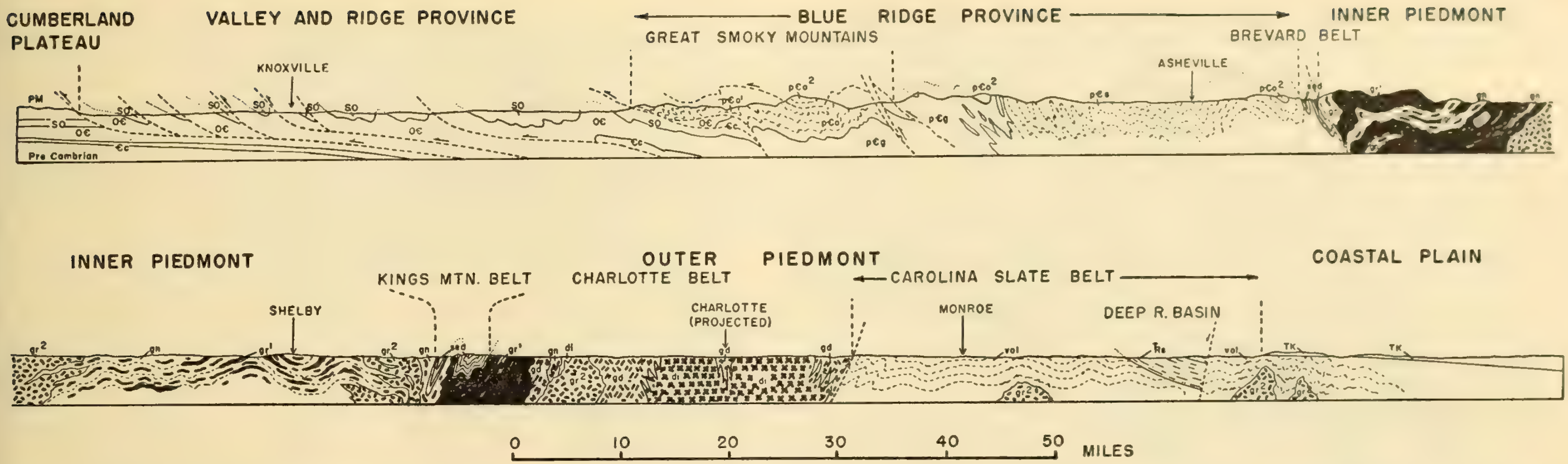


Fig. 7.2.. Cross section of Appalachian system from Cumberland Plateau to Atlantic Coastal Plain, from King, 1955 and 1959. Section B–B′, Fig. 7.1 Ŧs, Triassic Newark group; PM, Mississippian and Pennsylvanian rocks; SO, Middle and Upper Ordovician and Silurian rocks; O€, Cambrian and Lower Ordovician rocks; €c, basal Cambrian Chilhowee group; p€O^2, Great Smoky conglomerate and related rocks; p€o′, Hiwassee slate and Snobird fm.; p€s, gneiss and schist (mainly Carolina and Roan gneisses); p€g, Cranberry and Max Patch granites; vol, slate, tuff, rhyolite and andesite flows and breccia interbedded; gr^2, massive granites; gr′, foliated granites; di, diorite and gaboro; gd, granite-diorite injection complex; gn, gneiss and schist.

in the Atlantic Ocean to the margin of the continental shelf, so that the province geologically should be considered to include the continental shelf. It is clear that coastal plain sediments are being deposited today.

In addition to the great longitudinal structural divisions of the Atlantic margin of the continent just described, a traverse division is also commonly made, and the terms central Appalachians and southern Appalachians are used. Generally, the three structural systems, the folded and faulted Appalachians, the Blue Ridge, and the Piedmont provinces in the states of Alabama, Georgia, Tennessee, North Carolina, and Virginia are included in the southern Appalachian region, and the same three divisions in northern West Virginia, Maryland, Pennsylvania, and New Jersey are included in the central Appalachian region, although some authors call the whole structural complex south of New York the southern Appalachians.

RELATIONS TO GEOMORPHIC PROVINCES

Appalachian Plateaus Province

The structural divisions or systems are in large part reflected in the geomorphic provinces and, therefore, except for minor variations, their boundaries are the same. See Fig. 7.3. The Appalachian plateaus province includes two main plateaus, the Cumberland on the south, and the Allegheny on the north. The province is one of mature or submature dissection, and stands throughout four-fifths of its periphery higher than its neighbors; and parts of it are properly called mountains. The province is a broad, gentle, synclinal basin, whose youngest rocks are the Dunkard group or "upper barren measures" (Permian). They are mainly a thick mass of red shale and sandstone, and occupy a belt extending southwest from near Pittsburgh to near Huntington, West Virginia. Cropping out

Fig. 7.3. Block diagram of the geomorphic provinces of the central Appalachians and the Atlantic Coastal Plain, reproduced from Johnson, Bascom, and Sharp, 1933. M, Manhattan; Sb, Stroudsburg; P. Pottsville; R, Reading; Hb, Harrisburg; Cl, Carlisle G, Gettysburg; Ch, Chambersburg; Mr, Mercersburg; H, Hagerstown; HF, Harpers Ferry; F, Frederick; Rv, Rockville; Wash, Washington; Bal, Baltimore; Phil, Philadelphia; Tr, Trenton.

around it in successive elliptical zones are the Monongahela ("upper productive"), Conemaugh ("lower barren"), Allegheny ("lower productive"), and finally the fairly thin Pottsville. Most of the limestone and the best coal beds are in the Monongahela formation.

The Allegheny plateau is continuous with the Cumberland plateau, and any boundary is arbitrary. The southern plateau is somewhat less dissected, and the nearly flat-lying strata are largely the sandstones, shales, and basal conglomerates of the Pottsville formation.

. . . in southern Ohio the Mississippian rocks on the western margin of the Allegheny Plateau form cuestas rising to the full height of the plateau. The prominence of these cuestas diminishes toward the south, but they continue to form a narrow belt included in the plateau as far as latitude 37° 30′, beyond which the Mississippian rocks (all except the uppermost) spread widely to the west at a lower level and belong to a different province. Farther south the strong conglomerates or sandstones at the base of the Pottsville (Rockcastle group) underlie and support the margin of the plateau. All beds here dip slightly to the east, and the strong basal formations are to some extent stripped, leaving at places a decided eastward dip slope. As the stripped belt widens toward the south, and the province narrows, the entire width of the Cumberland Plateau in Tennessee and Alabama comes to be on the strong formations here known as Walden and Lookout sandstones.

For nearly 200 miles along the median line of the province in Tennessee and Alabama, runs the straight Sequatchie anticline, broken on the west by a thrust fault. If left uneroded, it would form a range of mountains, as it still does at its northern end where the Crab Orchard Mountains are in line with the perfect anticlinal valley which marks the rest of the uplift. Like the more extensive and complex Allegheny and Cumberland Mountains, this anticline represents the propagation into the plateau of the compressive stress by which the Valley and Ridge province was folded. Parallel to this feature, and 15 miles to the east is the similar Wills Creek anticline, marked by the valley west of Lookout Mountain (Fenneman, 1937).

Valley and Ridge Province

The folded and thrust-faulted Appalachian structural system is the geomorphic Valley and Ridge province, which as already stated consists of parallel or subparallel ridges and valleys of 1000 to 2000 feet local relief. It has been spoken of as the newer Appalachians in contradistinction to the older Appalachians which would include the Blue Ridge and Piedmont provinces.

The Valley and Ridge province can readily be divided longitudinally into a northwestern section, in which high ridges alternate with valleys of moderate width (the "Valley and Ridge" section), and a broad southeastern lowland section (the "Great Valley"). This division is more or less apparent throughout the length of the province.

Except for a short distance in New York, the entire northwestern boundary of the province is an erosional escarpment formed on gently dipping or horizontal sediments of the Appalachian Plateau. From southern Pennsylvania to Alabama, the southeastern boundary is formed by the resistant rocks of the Blue Ridge, towering above the Great Valley. This boundary is erosional in origin, weaker Paleozoic sediments having been stripped from the Precambrian surface (in some places from resistant Cambrian quartzites) on which they were deposited. In other localities the contact of weak Paleozoic sediments with resistant crystalline rocks takes place along a low-angle thrust fault, and erosion has lowered the sediments northwest of the fracture plane.

The rocks of the province are Paleozoic sediments ranging in age from Cambrian to Pennsylvanian. Their resistance to erosion varies greatly and has a very important effect upon the topography. The broad lowland composing the Great Valley is due to the weakness of the Cambro-Ordovician limestones (Kittatinny and other formations) and Ordovician shales (Martinsburg). The ridges of the Valley and Ridge belt are composed of very resistant middle and upper Paleozoic sandstones and conglomerates, particularly the Tuscarora quartzite and conglomerate (Silurian), the Pocono sandstone (Mississippian), and the Pottsville conglomerate (Pennsylvanian).

At the end of Paleozoic time the sediments in the Newer Appalachian province were subjected to strong pressure from the southeast and folded into great anticlines and synclines, in places overturned toward the northwest. Reverse faults were also commonly developed in the zone of greatest pressure, the horizontal attitude of the beds was scarcely disturbed. The region of undisturbed rocks today forms the Appalachian Plateau; the folded area has become the Newer Appalachians. In the latter province the structural trends are northeasterly, and owing to the remarkable development of subsequent streams the topographic features trend in the same direction (Fenneman, 1937).

Blue Ridge Province

The Blue Ridge province rises in southern Pennsylvania as the Carlisle prong and continues southwestward in accordance with the general trend of the Appalachian systems to northern Georgia. It stands conspicuously above the Great Valley section of the Valley and Ridge province on the northwest and the much lower Piedmont province on the southeast. The province takes its name from the Blue Ridge in Virginia, which is a relatively narrow mountainous ridge that extends from the Potomac River 200 miles southwestward to Roanoke. It has an altitude of about 1000 feet near the Potomac, but attains an elevation of more than 4000 feet to the southwest. Southwest of Roanoke, the Blue Ridge province is a rolling plateau, about 10 to 65 miles wide and with an average elevation of 3000 feet. Its bounding escarpments are 1000 to 2000 feet high. This part of the province includes the Great Smokies which are the highest land east of the Rockies. Mount Rogers, near the northwestern escarpment in Virginia has an altitude of 5719 feet, and Mount Mitchell in North Carolina has an elevation of 6711 feet.

The Blue Ridge geomorphic province terminates southward in northern Georgia, just north of Gainsville, where the Piedmont and the Valley and Ridge provinces seem to close around the Great Smokies. The Blue Ridge structural belt, however, extends on southward into Alabama, where it is buried by the coastal plain sediments; but because it has been eroded down to the level of the Piedmont, it is generally included in the Piedmont province by the geomorphologists.

The Piedmont province emerges from the Triassic lowlands in New Jersey, where it is known as the Trenton prong (see Fig. 7.3), and extends southwestward to Alabama. It is only a few miles wide in Pennsylvania, Maryland, and northern Virginia, but widens conspicuously to about 170 miles in North Carolina, from which place southwestward it continues wide. The surface of the Piedmont rises gradually westward to the foot of the Blue Ridge, where it reaches an altitude of 500 feet at the north and 1500 feet at the south. It is a vast plain along the horizon, but is maturely dissected to a local relief of a few hundred feet in places. Numerous hills and ridges rise as monadnocks 200 to 1000 feet above the general plains surface, and are more numerous near the Blue Ridge escarpment.

The rocks of the Piedmont province are mostly granites, gneisses, and schists, with some belts of marble and quartzite, partly of Paleozoic age but also in part of Precambrian age. A belt of basic rocks containing talc and soapstone is found near the western border. Several elongate basins of Upper Triassic sandstones and shales, cut by diabase dikes and sills, are found in the province. The Richmond basin contains coal, which was the first mined in North America in about 1750.

The Piedmont crystallines are overlapped on the east by the Cretaceous and Tertiary sediments, and the boundary of the two provinces is called the fall zone. Baltimore, Washington, Fredericksburg, Richmond, Petersburg, and other cities are located along it, and also mark approximately the points to which the tide extends up the estuaries.

8.

SOUTHERN AND CENTRAL APPALACHIANS

EXTENT AND DIVISIONS

The southern and central Appalachians extend from Alabama to New York and the Hudson River, and include the area shown on the index map of Fig. 7.1. They will be treated under their three longitudinal divisions, the folded and thrust-faulted Appalachian Mountains province, the Blue Ridge Cambrian and Precambrian province, and the Piedmont crystalline province. The use of the words southern and central implies that a northern division is also recognized, but this is referred to as the New England province. New Brunswick and Nova Scotia will be included in the northern division because of their close geological relation to New England.

MAJOR ELEMENTS OF STRATIGRAPHY

Appalachian Geosyncline

From the time that James Hall contributed voluminously to geologic literature (1840 to 1860) to about 1920, the following views were widely held regarding the Appalachian geosyncline. It extended from Newfoundland to Alabama and beyond, over 3000 miles; subsided most in the site of the present Valley and Ridge province and the eastern side of the Allegheny synclinorium, where more than 30,000 feet of sediments accumulated in places; shallow shelf seas extended inland from the geosyncline over the Central Stable Region; and a great borderland, Appalachia, lay along its southeast side, from which came much of the sediment that filled the subsiding trough.

Failure to appreciate facies changes and the absence of detailed mapping, especially in the Blue Ridge and Piedmont provinces, militated against a correct understanding of the tectonic development of the region. It appears now that the Blue Ridge province marks approximately the boundary between a west-lying miogeosyncline and an east-lying eugeosyncline in Cambrian time, but in post-Cambrian Paleozoic time the Blue Ridge and Piedmont were generally emergent. The concept of a borderland that extended beyond the present continental shelf into the Atlantic ocean is discredited.

Because of the metamorphosed nature of the strata in the Piedmont and the almost complete failure to find fossils in them, the work of unraveling their stratigraphy and structure has been slow. The stratigraphy of the Valley and Ridge province, however, has received a great deal of attention. It will be seen that geosynclinal subsidence in the site of the Appalachians and the plateaus shifted from time to time and place to place so that a strict coincidence of structural divisions and the sedimentary provinces does not exist. In a broad way, however, the western half of the miogeosyncline is undeformed or cast only into very gentle folds—it is structurally the Allegheny synclinorium and physiographically the Plateaus province—whereas the eastern half of the miogeosyncline is the folded and thrust-faulted province.

		East-central Tennessee (Chilhowee Mountain)	Northeastern Tennessee (Johnson, Carter and Unicoi Counties)	Northern Virginia (Elkton and Harpers Ferry areas)
Lower Cambrian		Shady dolomite (in Miller Cove)	Shady dolomite	Tomstown dolomite
Lower Cambrian	Chilhowee group	Hesse quartzite	Erwin quartzite	Antietam quartzite
Lower Cambrian	Chilhowee group	Murray shale	Erwin quartzite	Antietam quartzite
Lower Cambrian	Chilhowee group	Nebo quartzite	Erwin quartzite	Antietam quartzite
Lower Cambrian	Chilhowee group	Nichols shale	Hampton shale	Harpers shale
Lower Cambrian	Chilhowee group	Cochran conglomerate	Unicoi formation (with basalt flows 1000–1500 feet below top)	Weverton quartzite
Lower Cambrian	Chilhowee group	Cochran conglomerate	Unicoi formation (with basalt flows 1000–1500 feet below top)	Loudoun formation (with tuffaceous slate and rare flows)
Precambrian		Ocoee series	Volcanics of Mt. Rogers area; Cranberry granite	Injection complex

Fig. 8.1. Formations of the Chilhowee group in Tennessee and Virginia. From P. B. King, 1949.

Major Sedimentary Divisions of the Miogeosyncline

Lower Cambrian Marine Clastics. The oldest beds of the Cambrian, referred to as basal Cambrian, are conglomerates, arkoses, and shales, that pass upward into quartzites. They make up the Chilhowee group (Fig. 8.1) and attain a thickness of 5000 to 6000 feet. Tentative correlations with metamorphic units of the Piedmont suggest that these strata of the miogeosyncline grade southeasterly into eugeosynclinal facies in the manner illustrated in Fig. 8.2.

The basal Chilhowee beds rest in places unconformably on the volcanics and greenstones of the Ocoee series, and hence are believed to be part of the Lower Cambrian sequence. They are limited to a trough which runs the length of the central and southern Appalachians and are absent over the foreland or shelf region.

Cambrian and Lower Ordovician Carbonates. The miogeosyncline with its clastic deposits from Alabama to Pennsylvania became one dominantly of limestone and dolomite deposition. Some 9000 feet of carbonates representing the remainder of the Lower Cambrian, the entire Middle and Upper Cambrian, and the Lower Ordovician accumulated to a fairly uniform thickness up and down the entire trough. In the southern and northern ends of the geosyncline carbonate deposition continued into Middle Ordovician time. A correlation chart of the important formations of this period is given in Fig. 8.3. The carbonates grade into shale facies toward the northwest side of the miogeosyncline and the shelf in the manner illustrated in Fig. 8.4.

The basal Cambrian clastics and the succeeding thick carbonate sequence are typically miogeosynclinal and correspond in distribution approximately with the later orogenic belts of the Blue Ridge and Valley and Ridge provinces (King, 1959). The clastics were derived from an emergent stable interior, and the carbonates were deposited on a broad continental shelf, evidently without off-lying tectonic lands or a volcanic archipelago. The eugeosynclinal equivalents of the carbonates, if ever deposited, are not yet clearly recognized in the Piedmont.

Middle Ordovician Clastic Wedge. The regimen of erosion and sedimentation characterized by an emergent interior and a gently submerging continental border gave way abruptly in Middle Ordovician time to a reversed situation in which an uplifted borderland now furnished the sediments to a subsiding inside basin. The sediments were mostly clastic (Fig. 8.5), and the main source was in western Virginia, western North Carolina, and eastern Tennessee. A great fan of sediments is visualized to have apexed in this region in about the Great Smoky Mountains area and extended radially to the west, northwest, and north (P. B. King, 1959). See Fig. 8.31. It spread considerably beyond the later deformed belt of the Valley and Ridge province, and unlike the Cambrian and Lower Ordovician sediments was not confined to an elongate basin parallel with the continental margin. The wedge

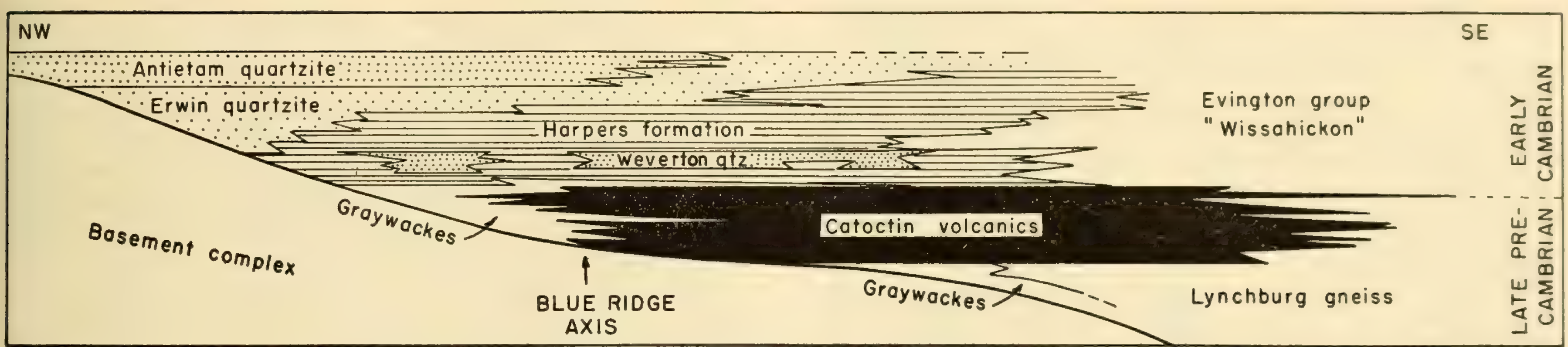


Fig. 8.2. Stratigrauhic relations of Late Precambrian and Early Cambrian formations in Blue Ridge of Virginia. After Bloomer and Werner, 1955.

or fan was about 8000 feet thick near its apex but thinned toward its edges. Beds representing the Middle Ordovician, as well as the Upper, are only 500 feet thick to the southwest in Alabama, and are carbonates. Likewise to the northeast in Pennsylvania Middle Ordovician beds are carbonates and only 700 feet thick.

Late Ordovician–Devonian Clastic Wedge. Apexing in east-central Pennsylvania is another great wedge of clastic sediments which began to accumulate in Late Ordovician time and continued through the Silurian and Devonian. The greatest subsidence and sediment accumulation occurred during the Late Devonian, which deposit is commonly referred to as the Catskill delta. It has a maximum thickness of over 8000 feet. Isopach maps of the Late Ordovician and Silurian deposits are shown in Fig. 8.6, and detail of facies relations in Fig. 8.7. A cross section of the Devonian wedge is given in Fig. 8.8, and a map of the deposit in Fig. 8.9. Further detail on the stratigraphy may be found in publications by Willard (1936).

Mississippian Deposits. In eastern Tennessee in the Great Valley a sheet of black shale may be seen transgressing across the Silurian strata and on the southeast side of the Valley to be resting on Middle Ordovician rocks. See Fig. 8.5. It is known as the Chattanooga shale and probably ranges in age from latest Devonian to earliest Mississippian (Rodgers, 1953). It thickens northeastward and eastward to a maximum of 400 feet at Cumberland gap. The Chattanooga shale is extremely widespread in the Nashville and Cincinnati arch areas and represents a marine facies of the upper continental beds of the Catskill delta.

The Mississippian above the Chattanooga in eastern Tennessee may have attained a maximum thickness of 6000 feet at the time of deposition near the Blue Ridge source region, but is generally much thinner than this in sections now preserved. It consists of three units each exhibiting a parallel gradation from finer, thinner, and less detrital—more carbonate sediments on the northwest side of the Great Valley to coarser, thicker, and more detrital sediments on the southeast side.

In Alabama, the thin Mississippian limestones of the foreland change toward the southeast into 5000 feet of sandstones and shales. In northern Virginia, Maryland, and Pennsylvania, the lower 1000 to 2000 feet of the Mississippian is shale and sandstone, the middle formations are limestone with a maximum thickness of 4000 to 5000 feet, and the upper formations are calcareous shale, red mudrock, and red and gray sandstones. The Pocono and similar sandstones of the lower division are thick bedded and conglomeratic. The thickening of most all units of the Mississippian from the western shelf to the eastern geosynclinal trough is conspicuous, and the coarsest material occurs where the section is thickest.

	TIME SCALE	CENTRAL PENNSYLVANIA	SUSQUEHANNA-NEW RIVER	NEW RIVER-TENNESSEE	TENNESSEE-ALABAMA
CANADIAN		LARKE	BEEKMANTOWN	CHEPULTEPEC	CHEPULTEPEC
UPPER CAMBRIAN	MADISON TREMPEALEAU	GATESBURG 1600-1750: MINES 150-200	CONOCOCHEAGUE 1600-2000	COPPER RIDGE 1200-2800	COPPER RIDGE 1200-2800
	FRANCONIA	GATESBURG 1600-1750: ORE HILL, STACY			BIBB 250-500 KETONA 400-600 BRIERFIELD 1500
	DRESBACH	WARRIOR 1250		NOLICHUCKY 400-750+	CONASAUGA: NOLICHUCKY
MIDDLE CAMBRIAN	MARJUM		ELBROOK 1800-3000	HONAKER: MARYVILLE 150-750	CONASAUGA: MARYVILLE
	WHEELER SWASEY	PLEASANT HILL 600		HONAKER: ROGERSVILLE 70-250	CONASAUGA: ROGERSVILLE
	DOME OPHIR HOWELL			HONAKER: RUTLEDGE 200-500	CONASAUGA: RUTLEDGE

Fig. 8.3. Middle and Upper Cambrian formations of central and southern Appalachians. After Resser, 1938.

The Mississippian trough coincides with the Valley and Ridge province and does not reflect the great westward bulging wedges of the Ordovician and Devonian. See Plate 6. It is probable that the Mississippian seas shored at about the Blue Ridge.

Mississippian rocks may never have been deposited in the northern part of the geosyncline in southeastern and eastern New York. The coarsest beds in eastern Pennsylvania were deposited nearest the highlands that formed in New England in the Devonian, and with reduction of the highlands the earlier Mississippian clastics were succeeded by calcareous sediments (Kay, 1942).

Pennsylvanian Clastics. The Pennsylvanian strata are distinctly clastic, both in the shelf and the geosynclinal areas. They are the great coal-bearing formations of the Allegheny Plateaus and Valley and Ridge provinces. A cross section from Virginia to Illinois that does not contain the present structural details is shown in Fig. 8.10. The trough is deepest in Alabama, where a maximum of 10,000 feet of strata—all Pottsville—is known. The Pottsville thins gradually northeastward until in Pennsylvania it is only 200 to 400 feet thick. As the Pottsville thins, younger Pennsylvanian formations appear, and in West Virginia and Pennsylvania the Allegheny formation is 300 feet thick, the Conemaugh 600 feet, and the Monongahela with the extremely valuable Pittsburgh coal at the base, 250 feet. The maximum thickness of the Upper Pennsylvanian is estimated to be 3000 feet.

The 10,000 feet of Pottsville beds in Alabama in the Coosa coal field area is rather restricted in east-west distribution because of the nearness of the Nashville arch to the Blue Ridge, but probably the original distribution was in the form of a wedge which spread westward over the site of the arch. This is the representation of King, 1959.

Permian System. Overlying the Monongahela formation in an oval area in West Virginia and Ohio, entirely in the Plateau province, is the Dunkard group or "upper barren measures" of Permian age. It is composed of shale, partly red, and sandstone with thin coal beds. Its maximum thickness is about 1500 feet.

FOLDED AND THRUST-FAULTED APPALACHIAN MOUNTAINS

Salients and Recesses

When viewed as a whole, the folded and thrust-faulted belt of the central and southern Appalachians consists of two major salients and three recesses. These are terms used by Keith (1923) in his well-known "Outlines of Appalachian structure." The salients are the arclike portions of the belt that are convex inland, and the recesses are the arclike portions

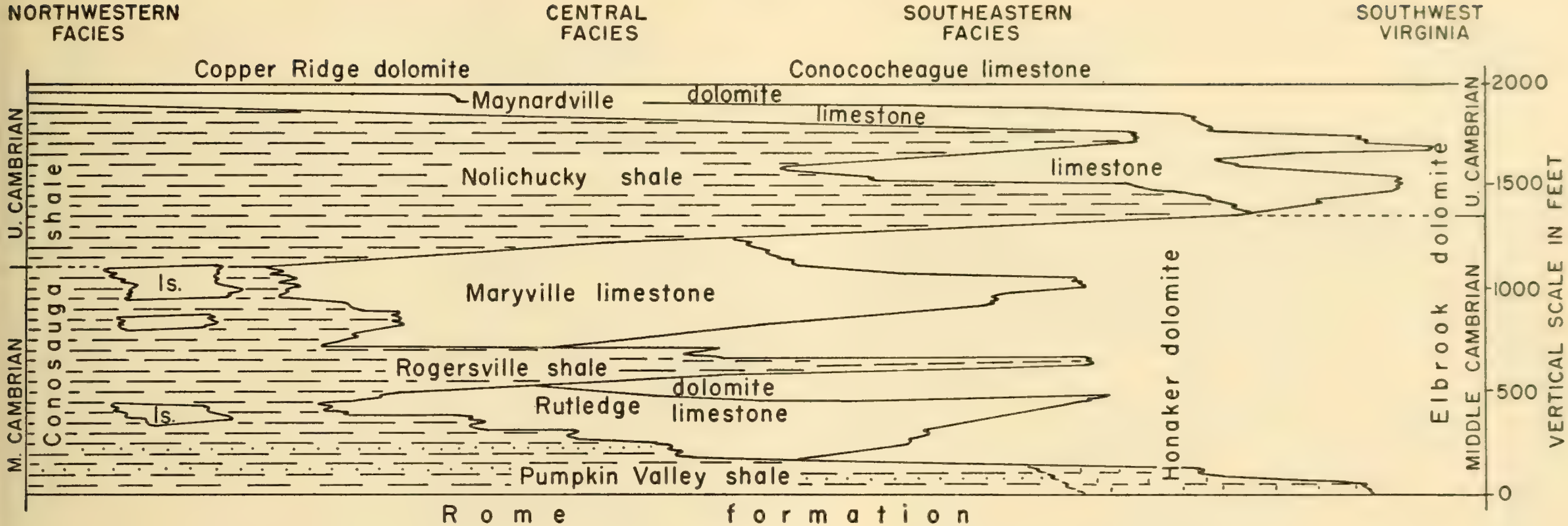


Fig. 8.4. Middle and Upper Cambrian sedimentary rocks of eastern Tennessee and southwestern Virginia. After Rogers, 1953.

that are convex toward the ocean. The southern salient is principally in Tennessee and southeastern Kentucky (see *Tectonic Map of the United States*), and the northern salient is in central Pennsylvania. They are about 400 miles apart. Keith points out two other salients in the northern Appalachians which will be described later.

Structural Characteristics in Alabama, Georgia, and Tennessee

If the *Tectonic Map of the United States* is studied, it will be seen that the southern half of the Valley and Ridge province is characterized by thrust faults, whereas the northern half is chiefly one of long parallel anticlines and synclines. In the southern part, the thrust sheets are stacked in imbricate fashion on top of each other, and in eastern Tennessee a succession of nine such sheets has been mapped. Some of the thrust sheets carry almost the entire Paleozoic succession; others duplicate the lower Paleozoic succession only. Precambrian rocks have nowhere in the belt been exposed as the result of thrusting and erosion.

The belt is made up almost entirely of thrust sheets in Tennessee, but southward, especially along the northwest margin, the beds are cast into a long anticline (Sequatchie) and syncline (Coalburg), which extend from central Tennessee almost to the Cretaceous cover in Alabama. Also along the southeast side of the belt in northwestern Georgia, a number of folds are evident. They occur in a conspicuous embayment of the Blue Ridge front.

The nature of the thrusts and folds is illustrated in sections 1 to 4 and 8 to 12 of Figs. 8.11 to 8.17. The location of the sections is given on the index map of Fig. 7.1. Most all the thrust sheets have moved toward the stable interior of the continent; only a few exceptions are known. One of these is illustrated in section 2, Fig. 8.12.

The Rome sheet was thrust forward at least 10 miles and then folded into anticlines and synclines. See section 3, Fig. 8.12. Some of the folds

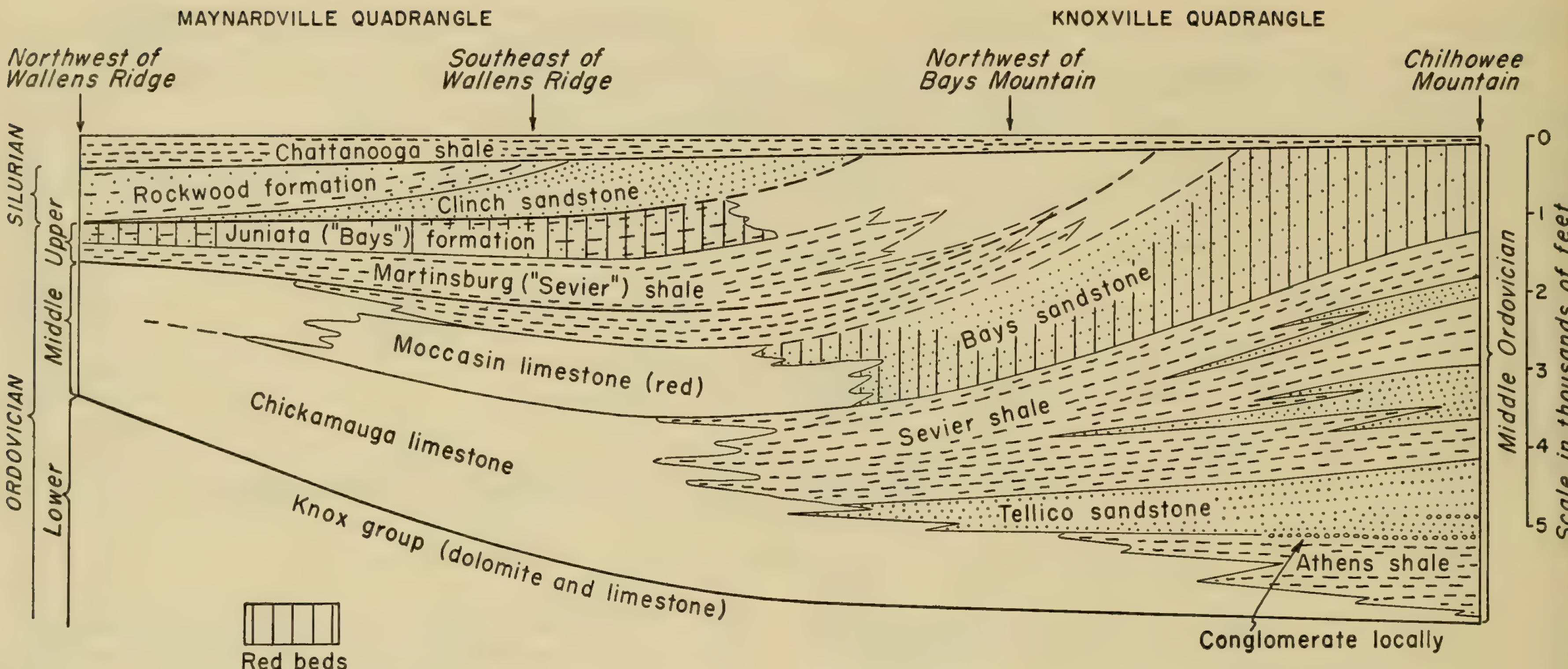


Fig. 8.5. Stratigraphic diagram of Middle and Upper Ordovician and Silurian rocks of Valley and Ridge province of eastern Tennessee. After P. B. King, 1950a.

of the strata below the thrust sheet are in the same position as those of the thrust sheet, but in detail the contacts are discontinuous against the thrust, and in other areas a complete lack of coincidence occurs. This suggests three episodes of compressional orogeny, perhaps almost in continuous succession: first the folding and erosion of the strata in front of the thrust and, perhaps, the early development of the thrust itself; then the movement of the great sheet out over the folded and eroded terrane; and third, further folding, involving both the thrust sheet and the underlying strata. Immediate waste products of the folds and thrusts which have been overridden and preserved, or which partially bury the structures, are not apparent. Such waste products in the form of coarse piedmont clastics are present in some of the Rocky Mountain thrusts and serve to date the various stages of deformation. Regarding the Rome thrust, however, all three closely related episodes of deformation are younger than the Lower Pennsylvanian Pottsville, which is involved in the deformation.

Another conspicuous structural division of the Valley and Ridge province of the southern Appalachians is the zone of shallow, flat thrust sheets, like the Rome, along its eastern margin. These are largely part of the Blue Ridge province, and involve Cambrian and Ordovician strata,

but in part are in the Great Valley. Modern interpretations show a number of fensters and klippes. See sections 4, 10, and 12 of Figs. 8.13, 8.16, and 8.17, respectively.

In northeastern Tennessee and southeastern Kentucky, the Appalachian front is characterized by an unusual thrust. Elsewhere the Appalachian front is one of fairly sharp folds that start abruptly from the flat-lying plateaus sediments. As seen in Figs. 8.14 and 8.15, an extensive block of the flat plateau strata has been torn loose and thrust, with only gentle deformation, toward the stable interior. The great, basal fault is known as the Pine Mountain and the two lateral tears as the Jackson and Russell Fork. Although the large mass is a thrust sheet, the strata from Pine Mountain to Cumberland Mountain are so flat that an arborescent drainage has developed and the region is considered geomorphically part of the plateaus province. The thrust mass is known as the Cumberland block and is 125 miles long and 25 miles wide. Its displacement has been calculated as 5.8 miles (Miller and Fuller, 1947). Along the Powell Valley anticline in the thrust sheet, erosion has cut several small fensters, and the Rose Hill oil field has been developed in the underlying beds with production from the Moccasin limestone.

Structural Characteristics in the Virginias, Maryland, and Pennsylvania

The southern part of the Appalachian belt, characterized by thrusting, is narrow; but toward the north in west-central Virginia a number of folds begin to show and the belt broadens. Sharp asymmetrical folds and mild metamorphism characterize the Great Valley, strong upright folds the main Valley and Ridge province, and very gentle folds, a western belt. See index map, Fig. 7.1. The folds of the westernmost zone are so gentle that the region is considered part of the Plateaus province, and the Appalachian structural front here is regarded as the western boundary of the zone of sharp folds. The plateaus generally stand in relief above the valleys and ridges of the strongly folded belt, and the eastward-facing escarpment is called the Allegheny front, which is a geomorphic feature, whereas the Appalachian front is a structural feature.

The chief faults are the Pulaski and North Mountain overthrusts. They may be parts of one great thrust which extends from southern Penn-

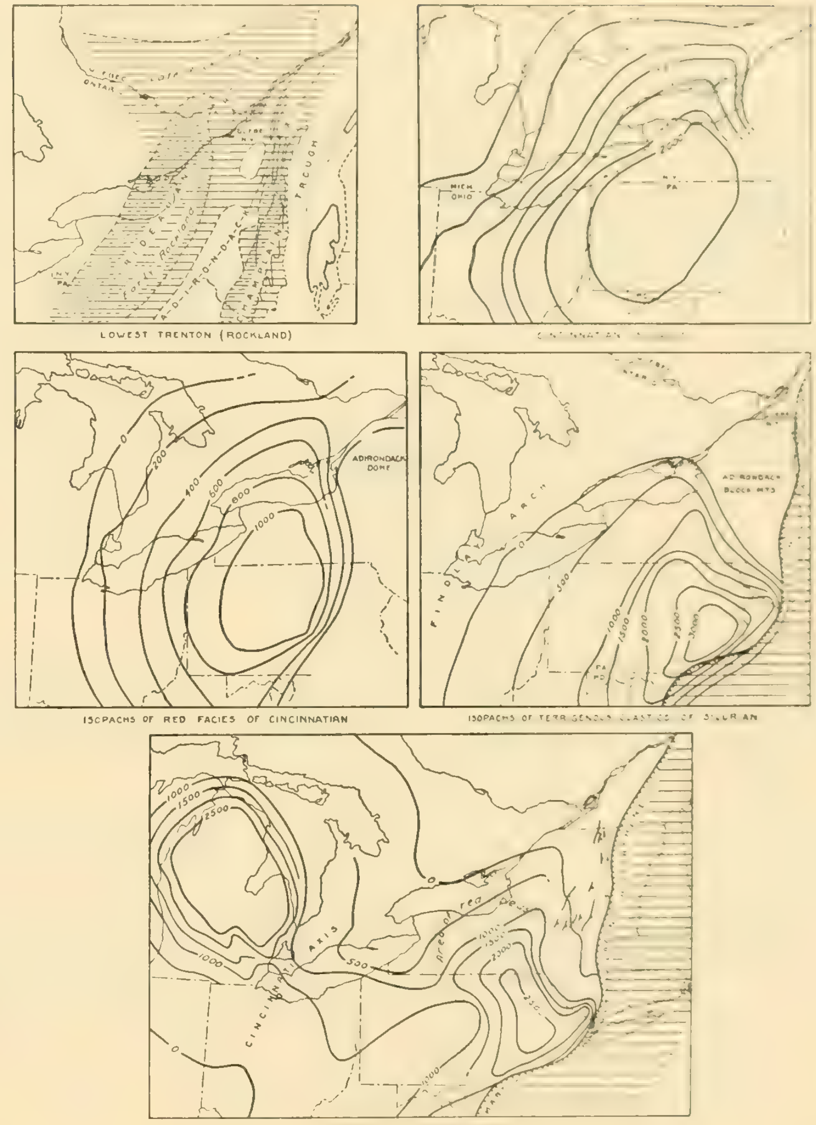


Fig. 8.6. Basins of deposition in middle and late Ordovician time and in Silurian time in the Pennsylvania–New York region. After Kay, 1942.

sylvania to northeastern Tennessee, over 500 miles long. Back of the Pulaski thrust front are several fensters, as illustrated in sections 12 and 13 of Fig. 8.17. See also Fig. 8.16. Sections 14, 16, and 17 of Figs. 8.18 and 8.19 also illustrate the thrusts of the central and eastern parts of the belt.

The nature of the strong folds is illustrated in sections 15, 16, 18–21, and 24 of Figs. 8.17 to 8.20. Most of the folds are asymmetrical and steepest on the northwest flank. According to the orthodox view, this marks active pressure from the southeast, as do almost all the thrusts. The folds of this region are some of the best known in North American geology, and some are markedly long and regular. See the *Tectonic Map of the United States.* Keith (1923) points out that the troughs of the folds extend downward to almost a common level, whereas the anticlines extend upward to variable elevations. Some of the anticlines are overturned and have broken into thrust faults. Most of the more eastern thrust sheets have extensively flat or folded lower surfaces.

The faults die out in southern Pennsylvania, and from there northwest-

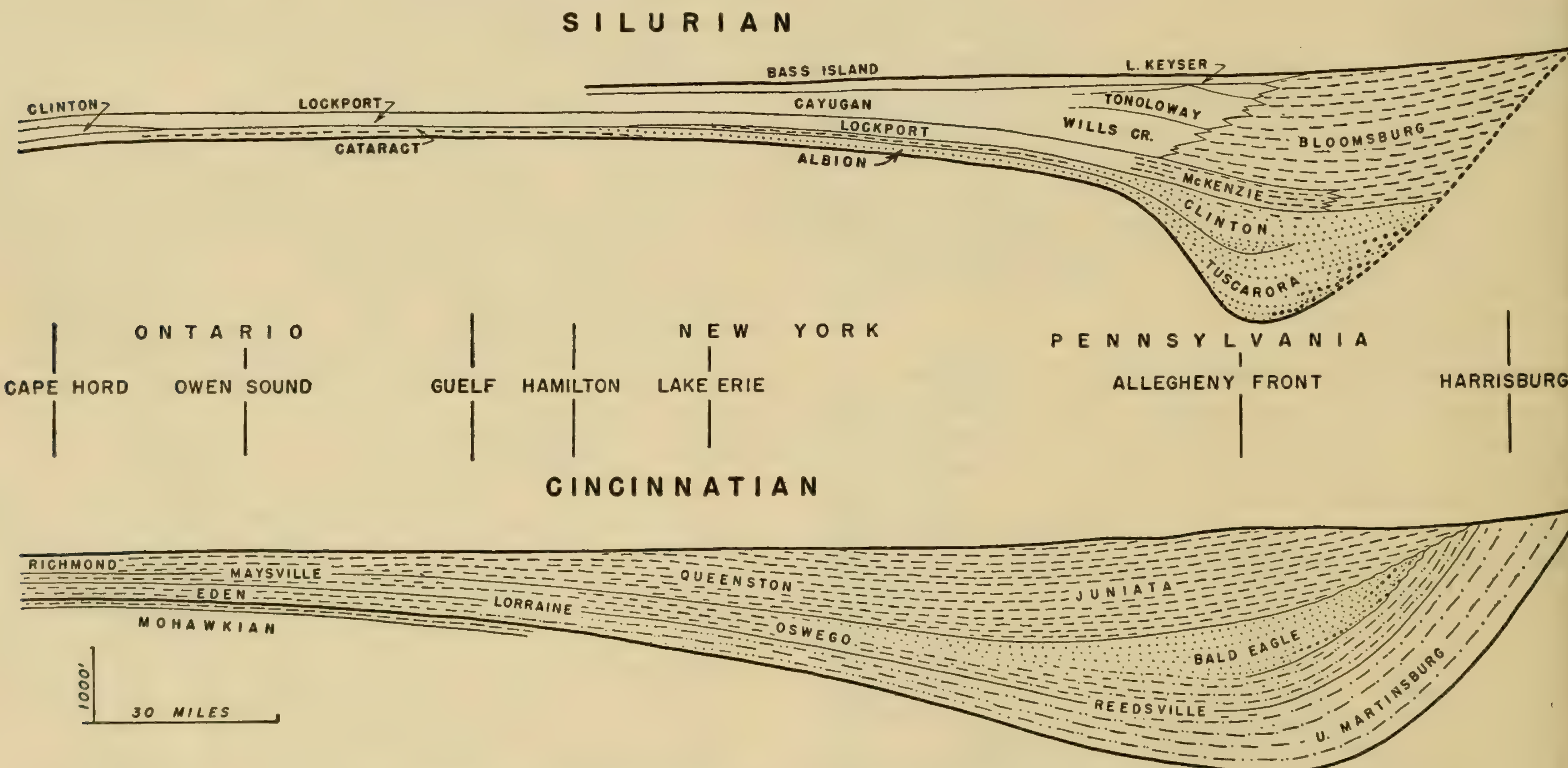


Fig. 8.7. Late Ordovician and Silurian stratigraphy of Pennsylvania, western New York, and western Ontario. After Kay, 1942.

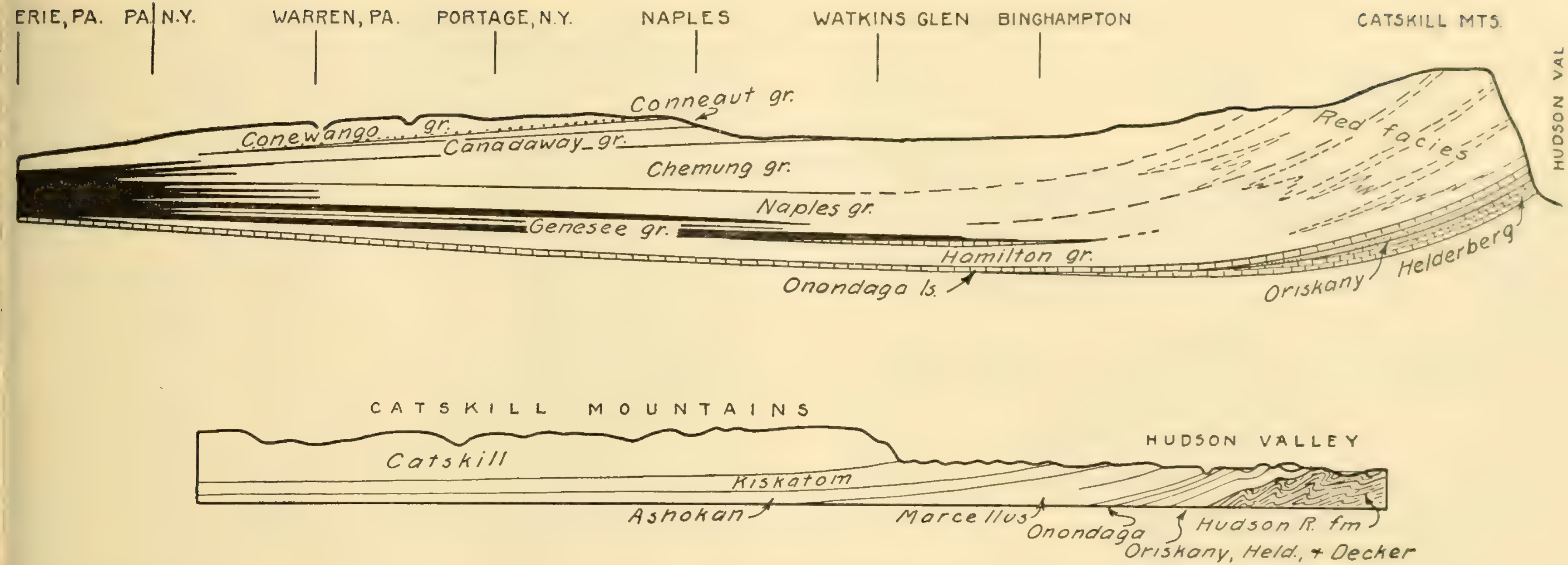


Fig. 8.8. Upper cross section, the great Catskill delta from Erie, Pa., to the Catskill Mountains, N.Y. After Schuchert, 1924.
Black is black shale and white is conglomerate, sandstone, shale, and calcareous shale. The clastics are dominantly red and generally coarsest in the eastern part. Vertical scale much exaggerated. Thickness may be judged by reference to the isopach map of Fig. 8.9.
Lower cross section, the Catskill Mountains and Hudson Valley north of Kingston, N. Y., after Chadwick and Kay, 1933. It shows the present eastern erosional termination of the Catskill delta, and presents the relations concerned with the problem of the source highlands.

ward almost the entire belt is one of anticlines and synclines. See section 29, Fig. 8.20. They veer markedly eastward in central and eastern Pennsylvania, and by southern New York both the gently folded belt and most of the strongly folded belt die out. The folds, if projected, would run into the Adirondack uplift and the lower Hudson Valley. A narrow eastern zone of the folded and thrust-faulted Appalachians, which is intimately connected with the Blue Ridge province, extends up the Hudson Valley. It seems very crowded between the Adirondacks and the New England metamorphic masses. See section 30, Fig. 8.21.

As far as the folded and thrust-faulted Appalachians are concerned, and aside from the narrow belt up the Hudson, it can be said that they begin in southern New York in gentle folds and become stronger southward. Thrust faults appear and become the dominant structure in the southern Appalachians. Also, in general, it can be said that the intensity of deformation across the belt becomes greater toward the southeast, and in the Great Valley and at the Blue Ridge front it is the greatest.

Regarding metamorphism, Keith (1923) pointed out long ago that a distinct change in constitution of the strata occurs along the eastern margin of the Valley and Ridge province in the Great Valley, and in the adjacent Blue Ridge. Shales have taken on a slaty character, limestones and dolomites are somewhat marmorized, and sandstones are quartzitic. The slate belt of northeastern Pennsylvania and southeastern New York in the tightly appressed and narrow belt of deformation east of the Blue Ridge is well known. The change from bituminous to semibituminous to

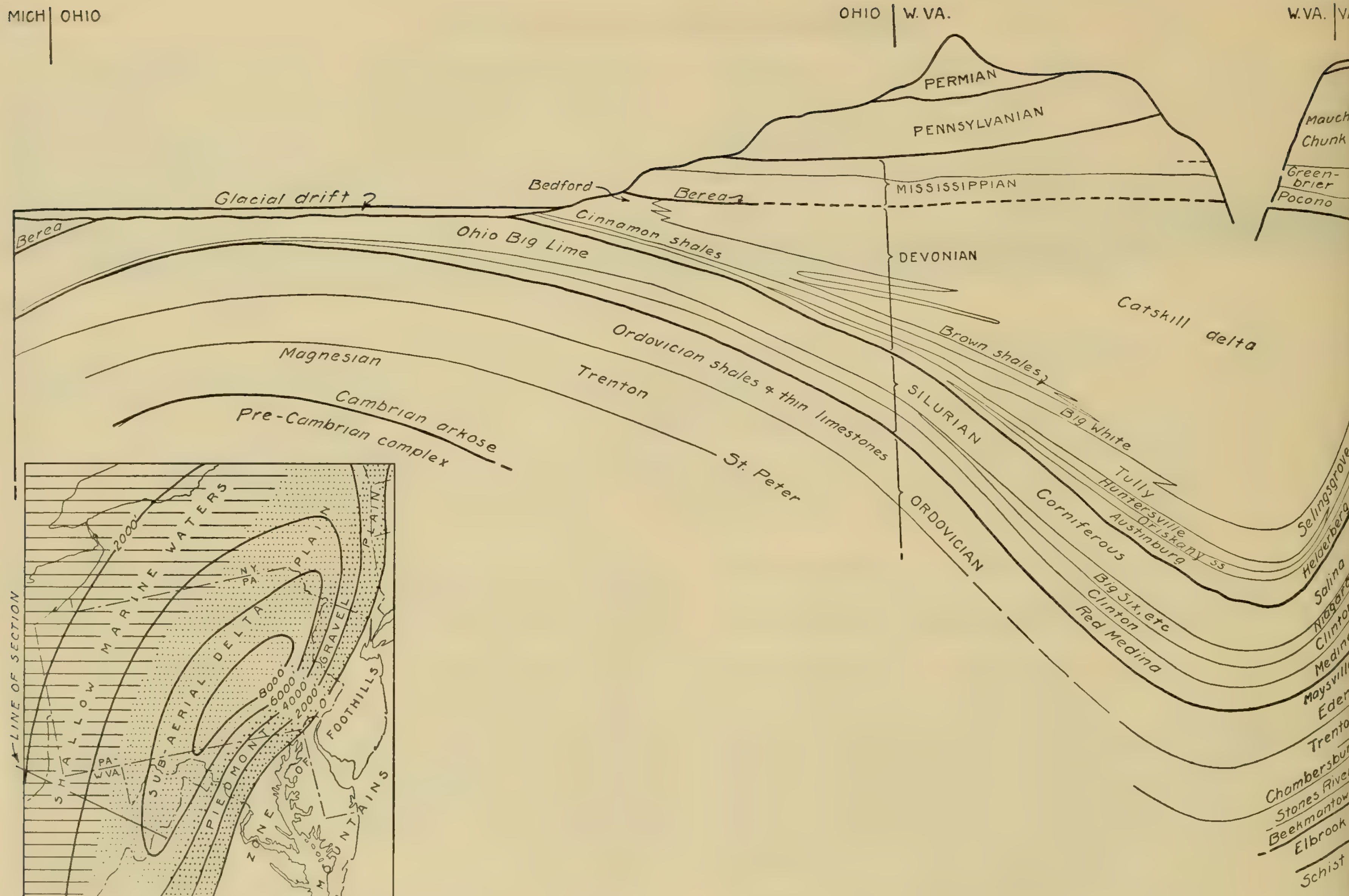


Fig. 8.9. Restored section of the Paleozoic rocks across the Allegheny basin and Cincinnati arch. The line of cross section is shown on the inset map, but it continues across Ohio to the southern Michigan line. After Tafferty, 1941, personal communication. The inset map shows the great Catskill delta and is taken from Barrell, in Schuchert's *Historical Geology*, 1924. The heavy lines are isopachs in feet.

anthracite coal eastward through Pennsylvania has been emphasized repeatedly as a demonstration of greater intensity of deformation from west to east. Although the coals have been metamorphosed within the belt west of the Great Valley, the associated shales, sandstones, and carbonates have not been much altered. Some doubts exist that the devolatilization is entirely a result of folding, because of anomalies in the relations, especially in West Virginia. Farther south the Knoxville, Tennessee, "marble" in the highly thrust-faulted belt is a slightly recrystallized rock.

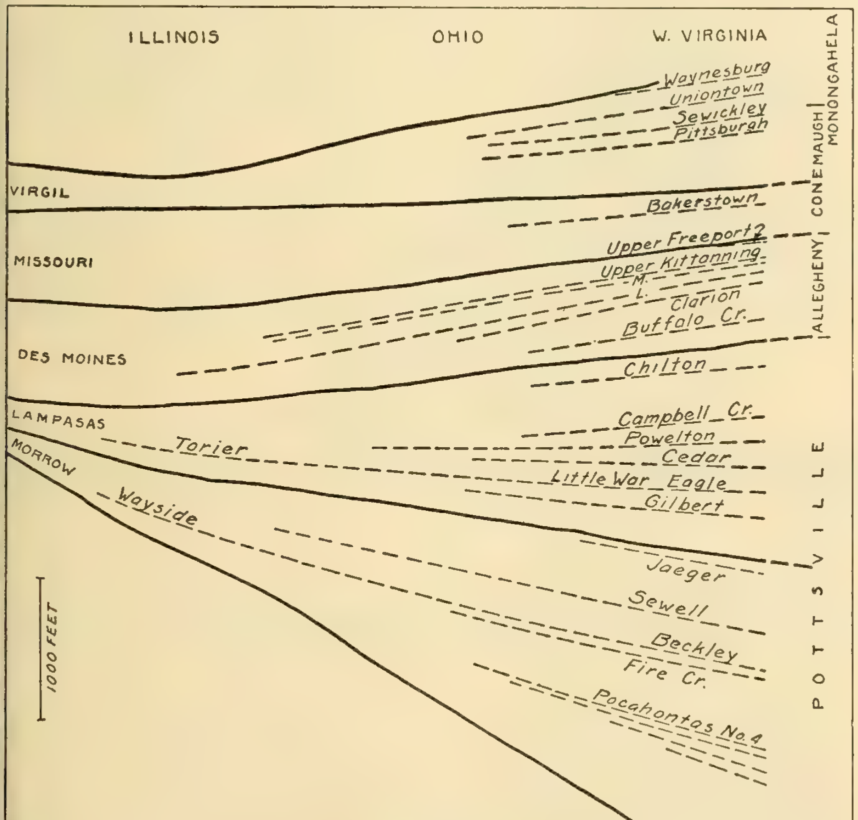


Fig. 8.10. Correlation and relative thickness diagram of Pennsylvania strata from West Virginia to Illinois. After committee report, *Chart No. 6 G.S.A.*, Vol. 55, 1944. The Cincinnati arch and other structures are not shown, nor is the section restored to any one time. The dashed lines are the various coal beds.

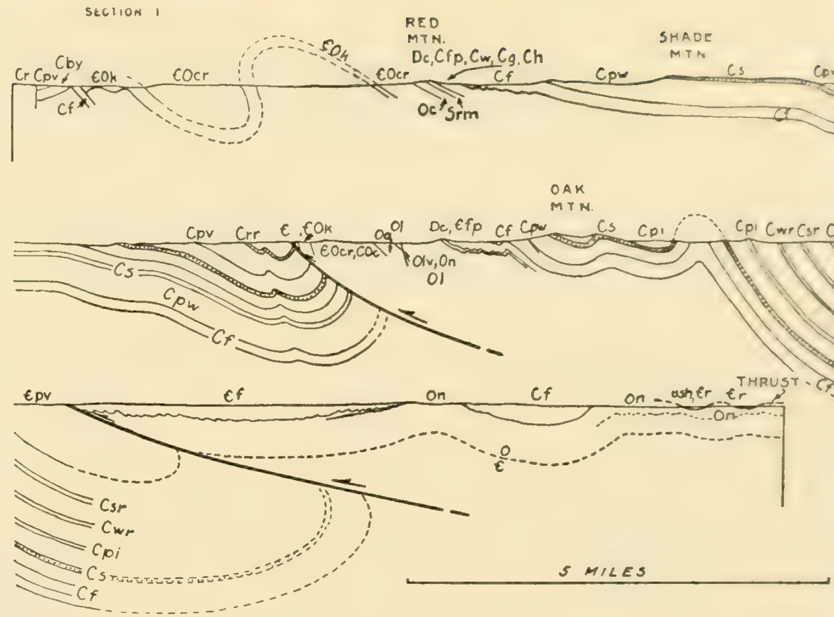


Fig. 8.11. Cross section (No. 1 of index map, Fig. 7.1) of the Bessemer and Vandiver quadrangles, Alabama, after Butts, 1927. Cr, Rome fm.; ЄOK, Ketona dol.; ЄOcr, Copper Ridge dol.; ЄOc, Chepultepec dol.; Olv, Longview ls.; On, Odenville, Newala, Lenoir and Mosheim lss.; Oa, Athens sh.; Ol, Little Oak ls.; Dc Chattanooga sh. and Frog Mtn. ss.; Cfp, Fort Payne chert; Cf, Floyd sh.; Cpw, Parkwood sh. and ss.; Cs, Cpv, Cpi, Cwr, Pottsville ss., sh., congl., and coal beds.

The change in constitution of the rock along the east side of the Great Valley is taken as a good boundary between the Valley and Ridge and Blue Ridge provinces by King (1950a).

Intrusive igneous rocks are almost entirely absent in the Valley and Ridge province, and hence no metamorphism incident to heat and volatiles is known.

BLUE RIDGE PROVINCE

Divisions

In the Blue Ridge and Piedmont provinces, we are confronted with a geology mostly of metamorphic and igneous rocks, only in part studied

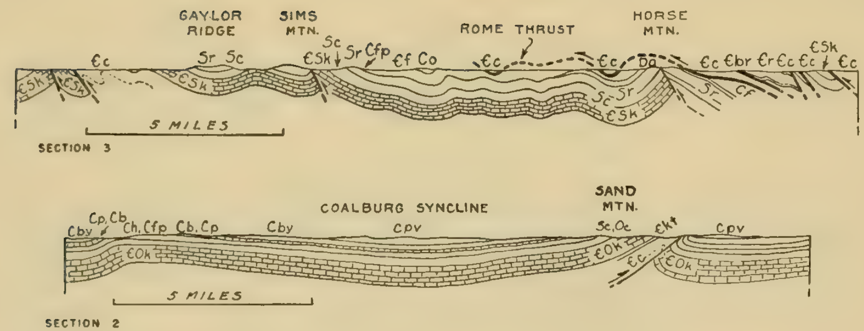


Fig. 8.12. Upper cross section (No. 3 of index map, Fig. 7.1) of Rome quadrangle, Georgia and Alabama. After Hayes, 1902.

Lower cross section (No. 2 of index map) of Birmingham quadrangle, Alabama. After Butts, 1910.

Єr, Rome fm.; Єc, Conasauga ls.; Єbr, Beaver ls.; ЄOk and Єsk, Knox dolomite; Oc, Chickamauga ls.; Sc. Clinton ss. and sh.; Sr, Rockwood fm.; Da, Armuchee chert; Cfp, Fort Payne chert; Ch, Floyd ss. member; Cb, Bangor ls.; Cp, Pennington sh.; Cby and Cpv, Pottsville gr.

and understood, and with an extensive literature that reveals a striking evolution of interpretation.

The Blue Ridge province embraces two rather distinct tectonic elements, about coincident with the geomorphic divisions. Northeastward from the vicinity of French Broad River in eastern Tennessee and western North Carolina, the Blue Ridge is narrow, whereas southwestward, it is broad and more complex. See Fig. 8.22.

Stratigraphy and Structure—Potomac to the French Broad River

The northeastern division, where most typically developed in northern Virginia, is composed of a core of older Precambrian crystalline basement rocks which are overlain and flanked by a considerable body of later Precambrian metavolcanics and metasediments (Catoctin greenstone and related units), and by Lower Cambrian clastic rocks (Chilhowee group). This segment is known as the Blue Ridge–Catoctin Mountain anticlinorium.

The structure and stratigraphy of the north end of the Blue Ridge belt of Fig. 8.22 across Catoctin Mountain and South Mountain is shown in Fig. 8.23. This sction is north of Harpers Ferry. Just south of the city the structure across Short Hill and the Blue Ridge is given in Fig. 8.24. Farther south in the Elkton area of Virginia a section on the west side of

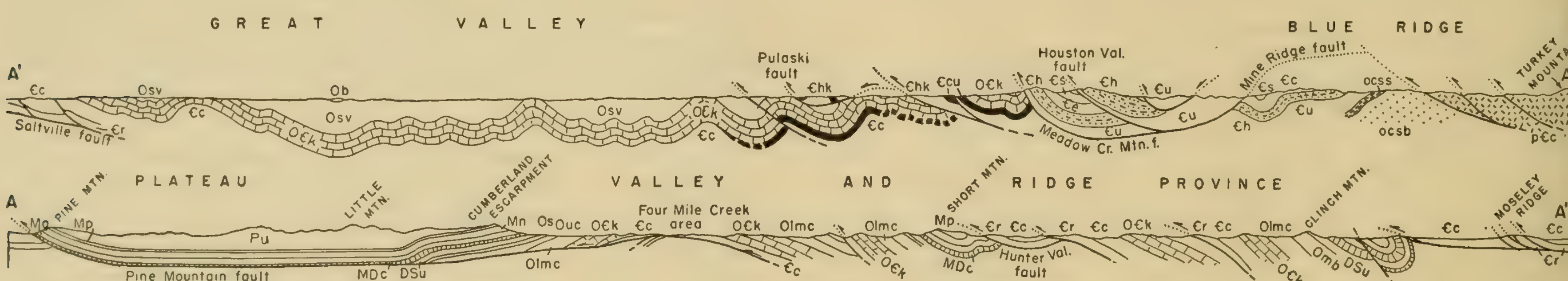


Fig. 8.13. Cross section of folded and thrust faulted Appalachians in eastern Tennessee. After Rodgers, 1953. (Section 4.) Pu, Pennsylvanian rocks; Mp, Pennington formation; Mn, Newman limestone; Mg, Grainger formation; MDu, Lower Mississippian and Upper Devonian rocks (Fort Payne, Grainger, and Chattanooga formations); MDc, Chattanooga shale; MDs, Basal Mississippian and Devonian shale; DSu, Lower Devonian and Silurian rocks (Hancock limestone, Rockwood formation, and Clinch sandstone); Os, Sequatchie formation; Oj, Juanita formation; Ouc, Upper part of Chickamauga limestone, including Reedsville shale; Omb, Martinsburg shale; Olmc, Lower and Middle parts of Chickamauga limestone, undivided; Ob, Bays formation; Oo, Ottosee shale; Oh, Holston formation, Ol, Lenoir limestone; Oa, Athens shale; Osv, Sevier shale; OЄk, Knox dolomite or group, undivided; Єc, Conasauga shale or group, undivided; Єcu, Upper Cambrian part of Conasauga group; Єhk, Monaker dolomite; Єr, Rome formation; Ss, Shady dolomite; Єe, Erwin formation and equivalent rocks (Hesse sandstone, Murray shale, and Nebo sandstone); Єhe, Hesse sandstone; Єnb, Nebo sandstone; Єh, Hampton formation; Єni, Nichols shale; Єu; Unicoi formation; Єch, Cochrane conglomerate; ocu, Ocoee series, undivided; ocss, Sandsuck shale; ocsb, Snowbird formation; pЄc, Precambrian crystalline complex.

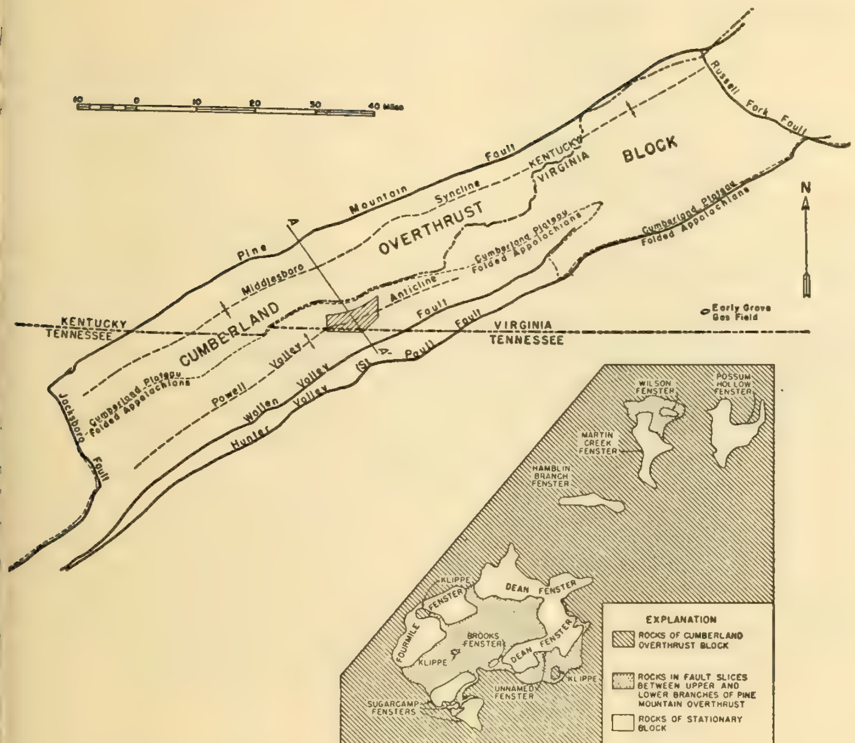


Fig. 8.14. Major structural features of the Cumberland overthrust block (upper map). Area of fensters ruled and shown in more detail in smaller map (lower). The area of fensters is now an oil field and is known as the Rose Hill district. Reproduced from Miller and Fuller, 1947.

the anticlinorial belt is as shown in Fig. 8.25. These recent interpretations of the structure show no important thrusts along the inner side of the Blue Ridge belt, but rather folded normal sequences. Southward, especially south of the James River, reverse faults are numerous and thrusting becomes dominant, as will be seen in the following discussion of the Great Smokies.

The old Precambrian crystalline complex is composed of granite, granodiorite, and gneiss. Cutting through it are basic dikes believed to be feeders of the overlying basaltic Catoctin greenstone. The whole Catoctin mass has undergone low-grade metamorphism, and a slaty or schistose cleavage pervades it, which dips southeastward as shown in the sections just referred to. A distinct lineation occurs along the general boundary of the Precambrian and Paleozoic rocks, and in northern Virginia, Maryland, and Pennsylvania, the cleavage in which the lineation lies extends into the Beekmantown beds according to Cloos (1957) and into the Martinsburg shale according to Nickelsen (1956). See Fig. 8.24. Lineation and cleavage is limited to the Precambrian from Roanoke southwestward where thrusting has brought the basement rocks into abrupt contact with the unaltered Paleozoic rocks.

The shear type of deformation accompanied by thickening along the fold axes and thinning along the flanks is most characteristic of the Blue Ridge, and sets it apart from the Valley and Ridge structures.

Only one deformation has been detected from the lineation north of the Potomac River in the South Mountain anticlinorium. Since the Precambrian Catoctin greenstone as well as the Cambro–Ordovician limestones and shales of the Great Valley are affected and since the lineation is remarkably regular along the Blue Ridge from Pennsylvania to the French Broad River in North Carolina, Cloos (1957) thinks that this one deformation is post-Ordovician, and therefore either Taconian or Acadian in age. These orogenies will be described presently.

Great Smoky Mountains

South of the French Broad River the Blue Ridge belt loses its weltlike form, and a broad, high, and geologically complex terrane sets in. Along the Tennessee–North Carolina boundary between the cities of Knoxville and Ashville are the Great Smoky Mountains where 16 peaks rise above 6000 feet. The general expansion of the Blue Ridge in this region is shown on Fig. 8.22, and a geological map by P. B. King is presented in Fig. 8.26. A small-scale cross section is part of Fig. 7.2, and a more detailed section is given in Fig. 8.27. Most of the Great Smokies is a thrust complex of the Ocoee Late Precambrian series.

> This is a body of terrigenous clastic sedimentary rocks, which has minor intercalations of limestone and dolomite but no volcanic components or known fossils. The series is probably 30,000 feet or more thick. It lies unconformably on a basement of earlier Precambrian granitic and gneissic rocks, and on the

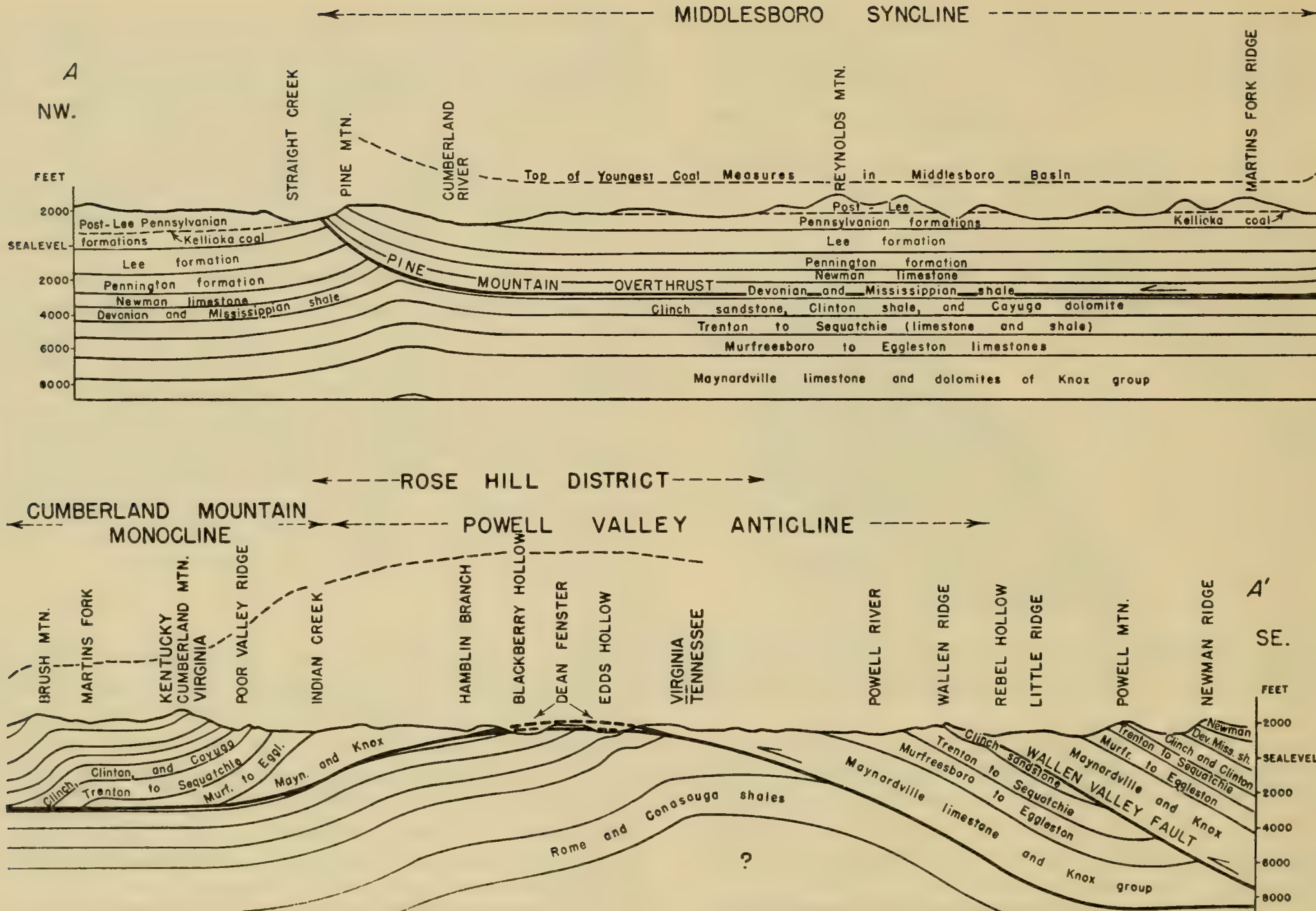


Fig. 8.15. Section across Cumberland overthrust block along line A–A′ of Fig. 8.14. Length of section, 27 miles. Displacement along Pine Mountain fault, 5.8 miles. Reproduced from Miller and Fuller, 1947. Section line A–A′, is line 8 on index map of Fig. 7.1.

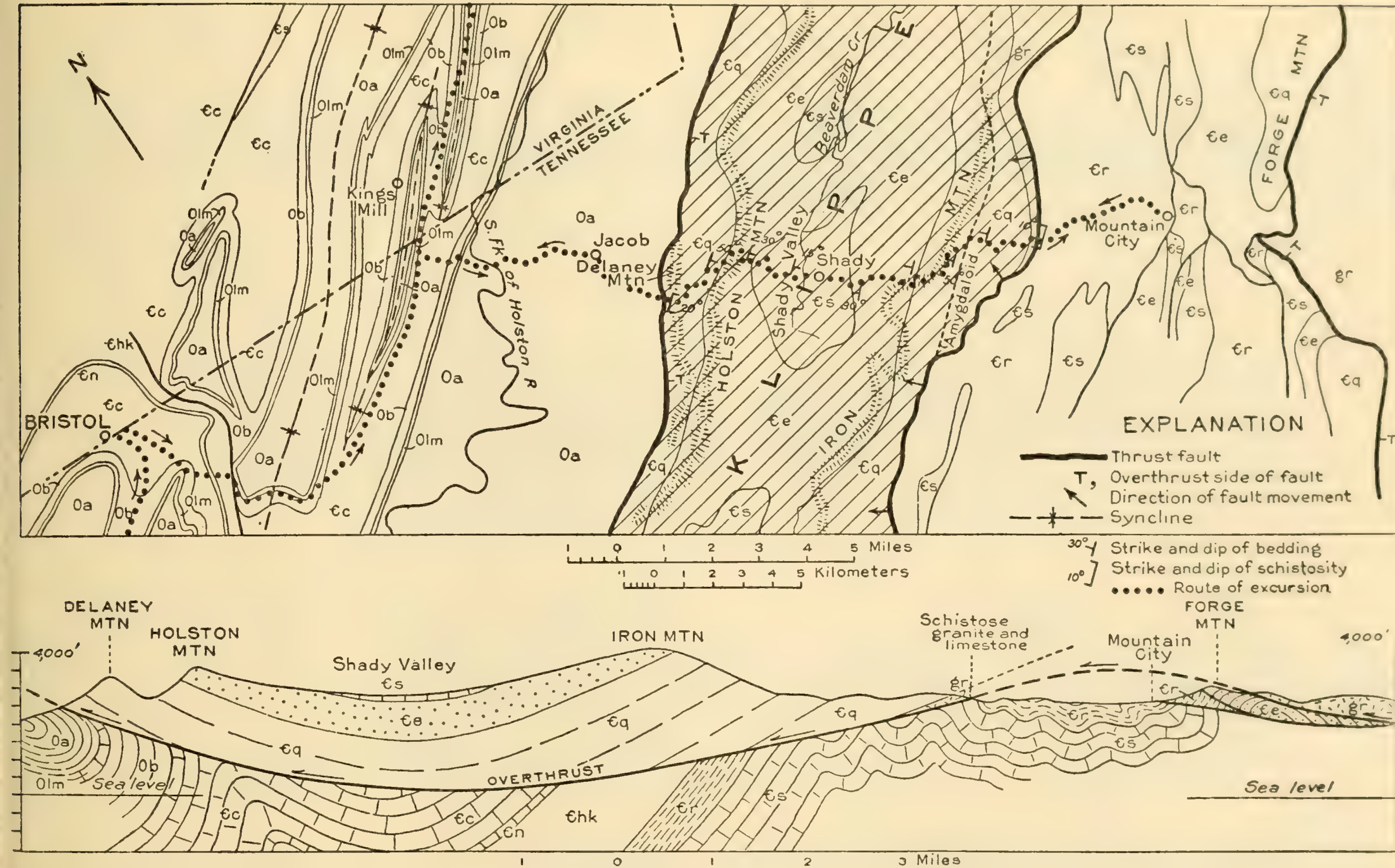


Fig. 8.16. Geologic map and section of the area from Bristol, Va., to Mountain City, Tenn. Reproduced from Butts *et al.*, 1933. Section 10 of index map, Fig. 7.1. Oa, Athens shale; Olm, Lenoir, and Mosheim limestones; Ob, Beekmantown dolomite (Nittany and post-Nittany); Cc, Conococheague limestone; Cn, Nolichucky shale; Chk, Honaker dolomite; Cr, Rome formation; Cs, Shady dolomite; Ce, Erwin quartzite; Cq, Cambrian quartzite and shale, undifferentiated; gr, Precambrian granite.

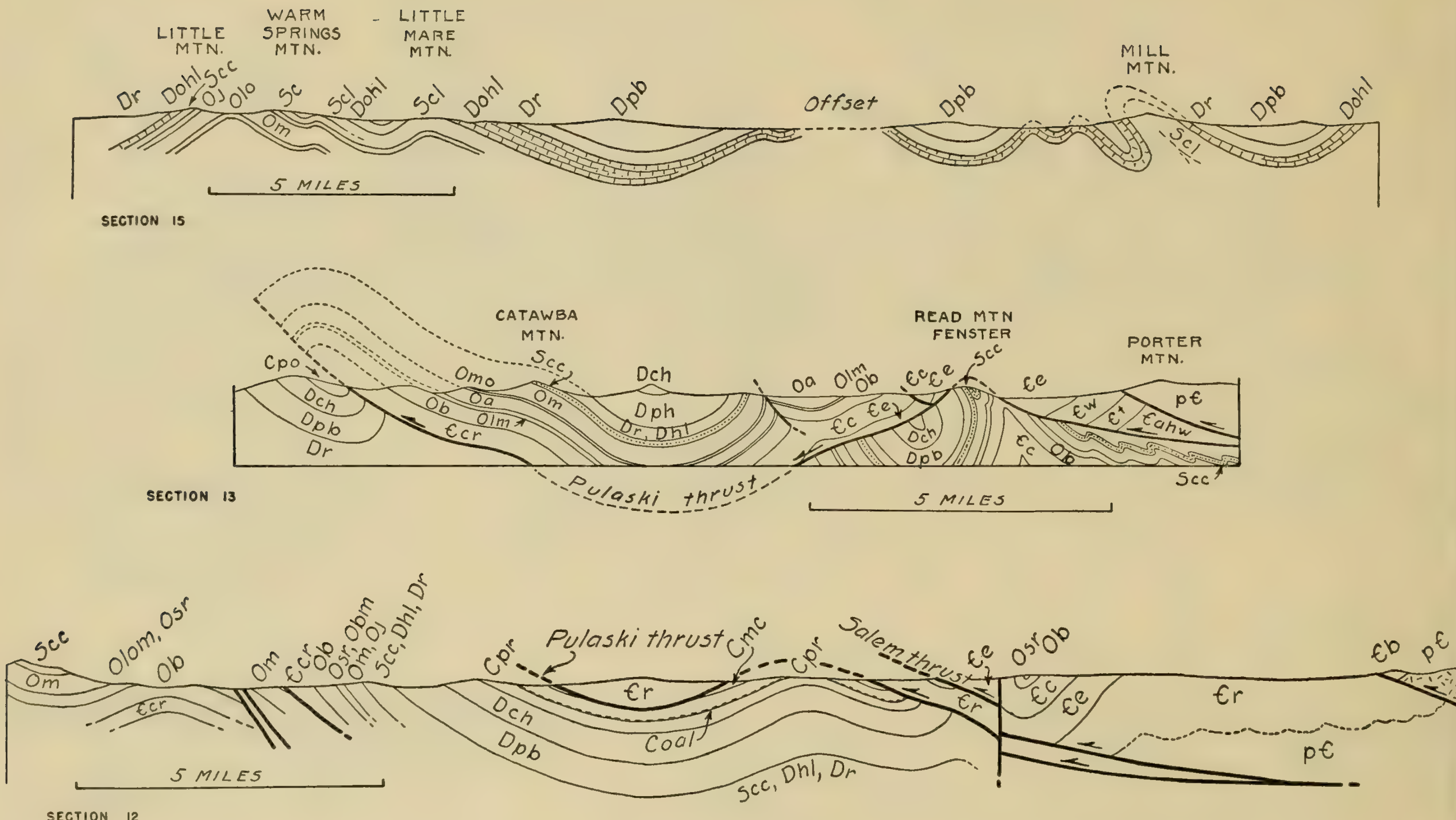


Fig. 8.17. Sections in the folded and thrust-faulted Appalachians of western Virginia, after Butts *et al.*, 1933. Section 15 is from Warm Springs to Goshen; Section 13 is through Hollins College; and Section 12 is through Newport and Christiansburg. See index map, Fig. 7.1.

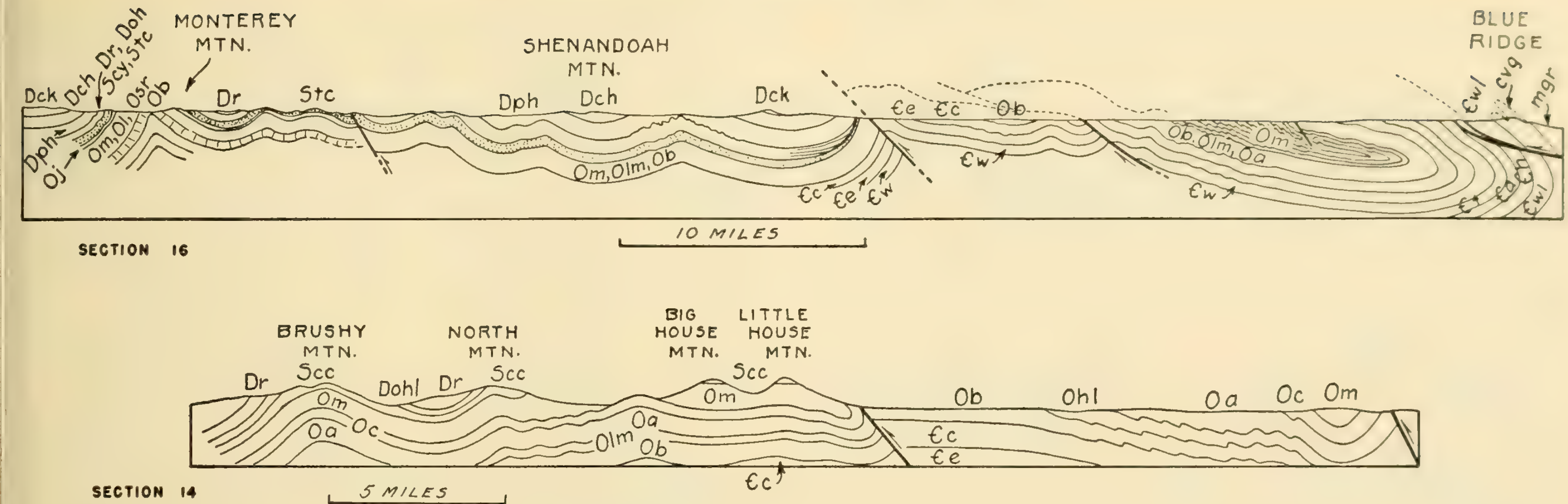


Fig. 8.18. Section in the folded and thrust-faulted Appalachians of west-central Virginia, after Butts *et al.*, 1933. Section 14 runs through Lexington and Section 16 from the West Virginia line to Waynesboro and Afton. See index map, Fig. 7.1.

northwest side of the mountains it is overlain by the Cochran formation, or basal unit of the Chilhowee group, which is of Cambrian and Precambrian (?) age. South of the mountains it is overlain by rocks of the Murphy marble belt; here, the top of the Ocoee is placed tentatively at the base of the Nantahala slate.

The Ocoee series is divisible into three broad units of regional extent and contrasting lithologic character, which are herewith designated groups and named the Snowbird group, the Great Smoky group, and the Walden Creek group. The groups consist of local intergrading and intertonguing formations and have complex stratigraphic and structural relations. The Ocoee series is split by major thrust faults into three sequences, a southern, central, and northern, none of which contains more than two groups of the series (King *et al.*, 1958).

The west front of the Great Smokies and the Blue Ridge belt southwestwardly is characterized by great folded thrusts, described in part under the previous Valley and Ridge province. Where the overridden rocks are exposed as re-entrants or windows and composed of limestone or dolomite, they form "coves" or valleys lying within the mountains of

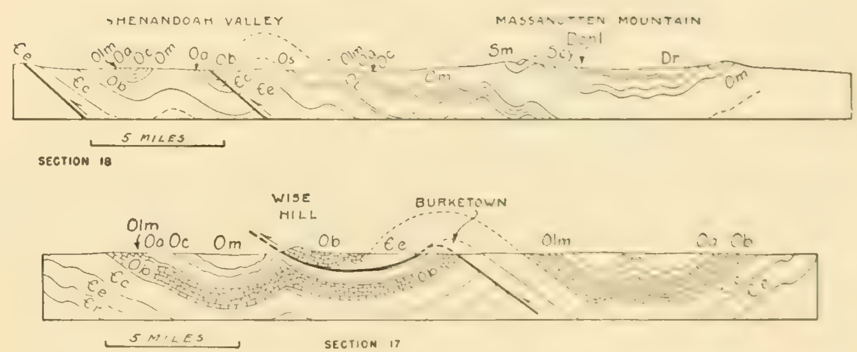


Fig. 8.19. Sections in the folded and thrust-faulted Appalachians of Virginia. After Butts *et al.*, 1933. Section 17 is about 10 miles north of Staunton, and Section 18 is 10 miles north of Shenandoah Caverns. See index map, Fig. 7.1.

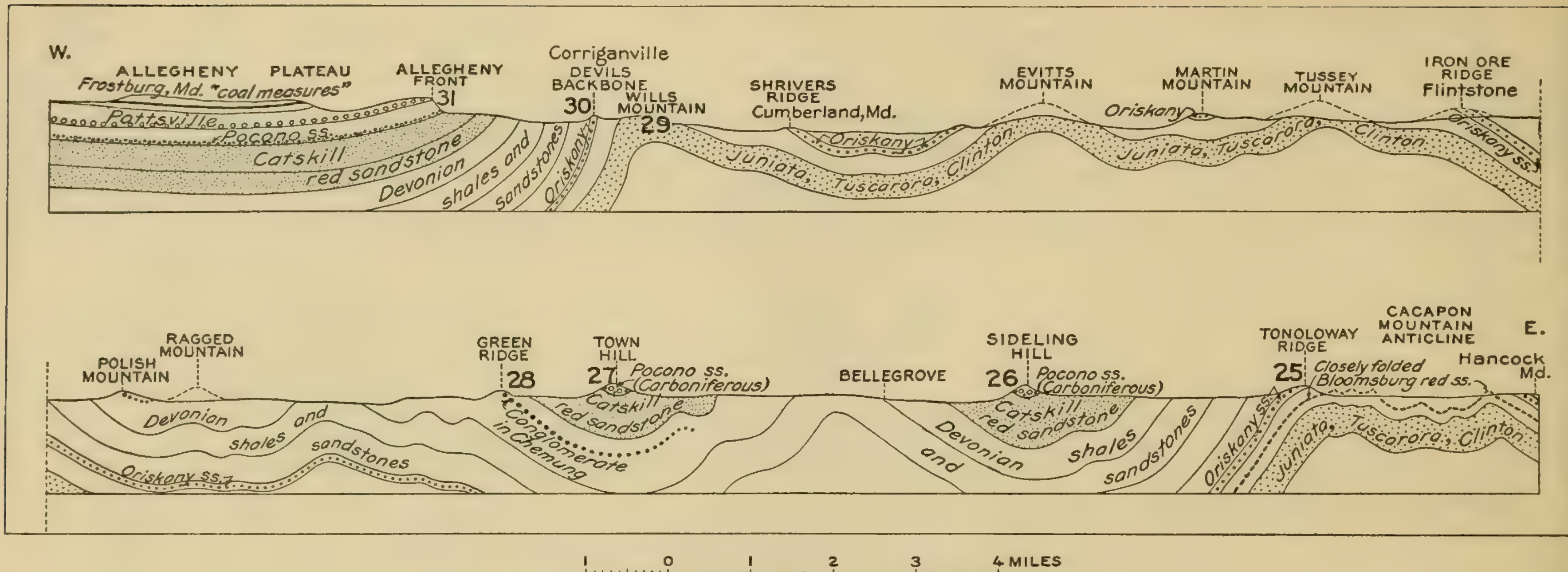


Fig. 8.20. Section from Allegheny plateau to Cacapon Mountain anticline at Hancock, Md. Reproduced from Butts *et al.*, 1933. Section 21 of index map, Fig. 7.1.

granitic or clastic rocks. One of the windows, such as that containing Grandfather Mountain, North Carolina, lies 35 miles southeast of the northwest boundary of the Blue Ridge province (King, 1951).

PIEDMONT PROVINCE

Principal Foliate Rocks

The most voluminous rocks in the crystalline Piedmont of Pennsylvania, Maryland, and Virginia is a great schist series which exhibits in part high-grade metamorphism (Jonas, 1932). It contains in places various metavolcanics. In Cecil and Harford counties, Maryland, a volcanic sequence well described by Marshall (1937) shows considerable variation, grading from massive amygdaloids and even-textured volcanics through schistose amygdaloids to fine-grained hornblende schists which in places are indistinguishable from sheared gabbro except by microscopic examination. Several bodies of mylonitized granite in the schist series have been recognized.

The most troublesome and yet unsolved problem is the age of the schist. It has been correlated with the Glenarm series of Pennsylvania and Maryland, and to this most authorities agree; but the age of the Glenarm is not yet known. It is generally believed to be late Precambrian or early Paleozoic.

A few anticlines and domes of older rocks, the Baltimore gneiss, are found within the schist series, in Maryland and Pennsylvania, and presumably others occur in Virginia.

The Piedmont and eastern part of the Blue Ridge in North and South

Carolina consists of a complex of contorted gneisses, containing granite plutons and satellitic offshoots, swarms of small ultrabasic intrusives, and narrow zones of metasedimentary rocks. The boundary of the Piedmont and Blue Ridge provinces is here indistinct on the basis of bedrock geology. The dominant unit of this complex is the Carolina gneiss. It consists of quartz, feldspar, mica, gneiss, and hornblende gneiss and these are considered to be originally sedimentary and volcanic rocks but altered incident to the batholithic intrusions. King (1951) points out that no clear break exists between the gneiss complex and the Ocoee and Talladega series in the Great Smokies to the northwest, and a considerable part of it may be a highly altered phase of these late Precambrian geosynclinal deposits.

Metamorphism

As noted by King (1951, 1959) the metamorphism increases progressively southeastward from the Great Valley across the Blue Ridge, into the Piedmont province, and climaxes with the development of silimanite in the central part of the Piedmont between the Brevard and Kings Mountain belts (see Fig. 7.2). Southeast of the silimanite zone the metamorphism is less intense. The belt of decreased metamorphism is marked

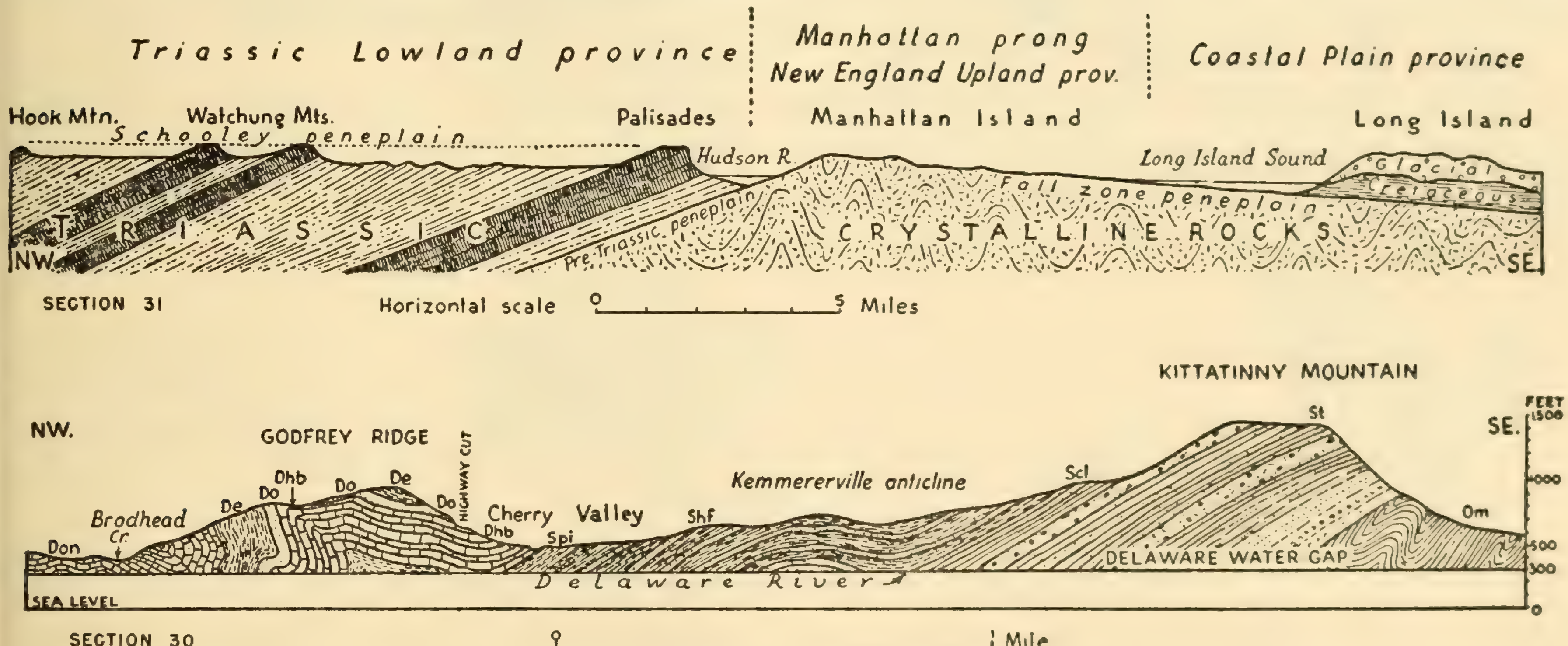


Fig. 8.21. Upper section across Triassic basin and Manhattan Island to Long Island, N. Y. Reproduced from Johnson *et al.*, 1933. No. 31, Fig. 7.1. Lower section through Delaware Water Gap. Johnson *et al.*, 1933, after Willard. Om, Martinsburg shale; St. Tuscarora ss.; Scl, Clinton ss.; Shf, High Falls sh.; Spi, Paxono Island sh.; Dhb, Helderberg ls.; Do, Onondago ss.; De, Esopuc grit; Don, Onondago ls.; Dm, Marcellus shale; Dh, Hamilton ss. No. 26, Fig. 7.1.

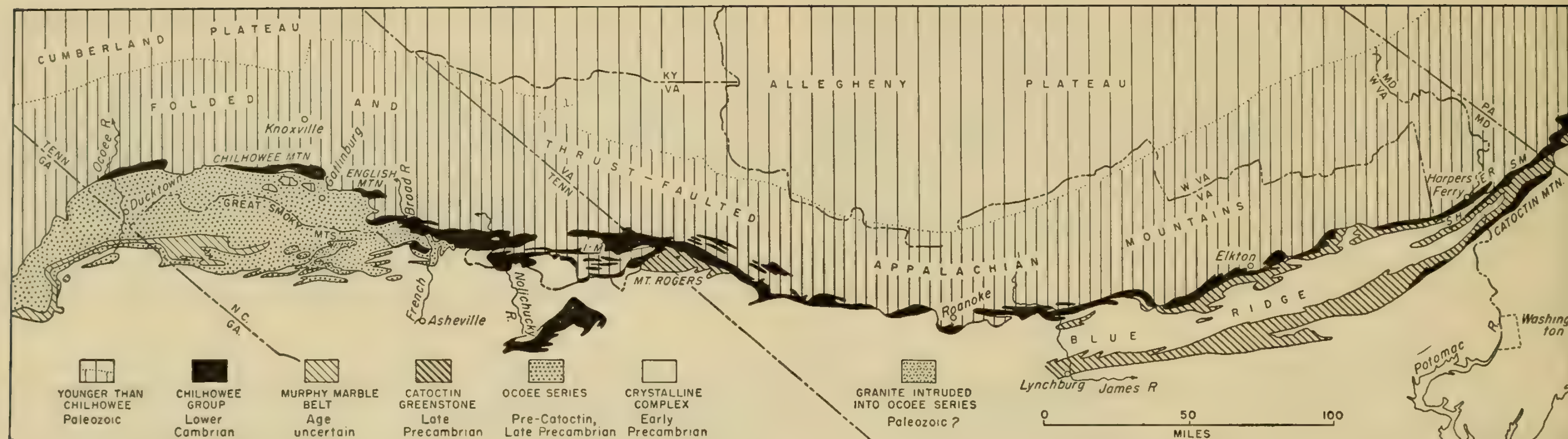


Fig. 8.22. The Blue Ridge province from Georgia to Pennsylvania showing principally the Lower Cambrian clastic group (Chilhowee) and the Late Precambrian Catoctin greenstone and Ocoee series The Catoctin greenstone includes volcanics and sediments of Mt. Rogers area. After P. B. King (1949). E.R., Elk Ridge; S.M., South Mountain; S.H., Short Hill; I.M., Iron Mountain

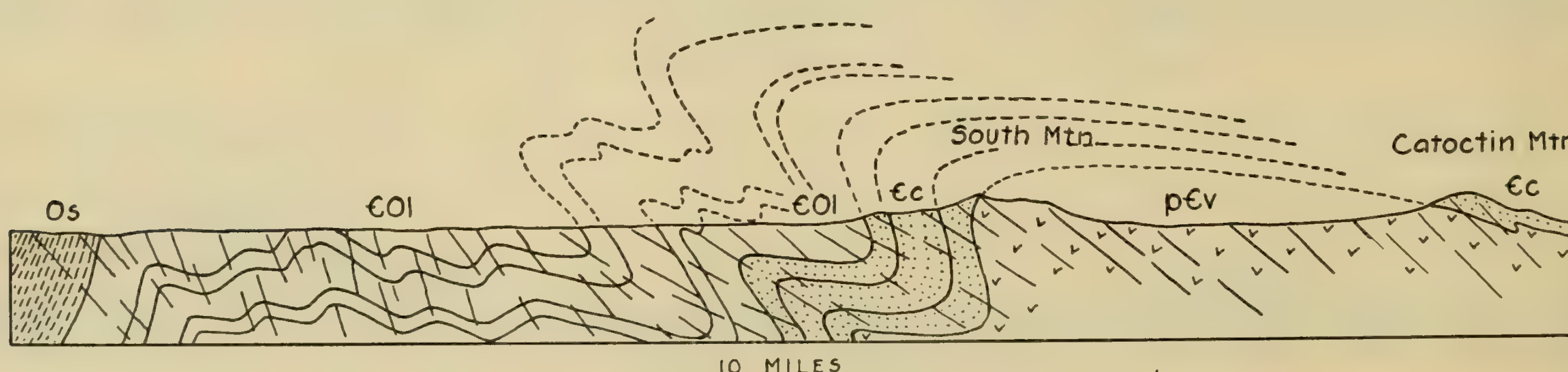


Fig. 8.23. Change from open to close folding along east side of Great Valley, in vicinity of South Mountain, Md. After P. B. King, 1950a. p€v, volcanic rocks; €c, Lower Cambrian Chilhowee gr.; €Ol, Cambrian and Ordovician ls., dol., and some sh.; Ordovician shade. See Fig. 8.22.

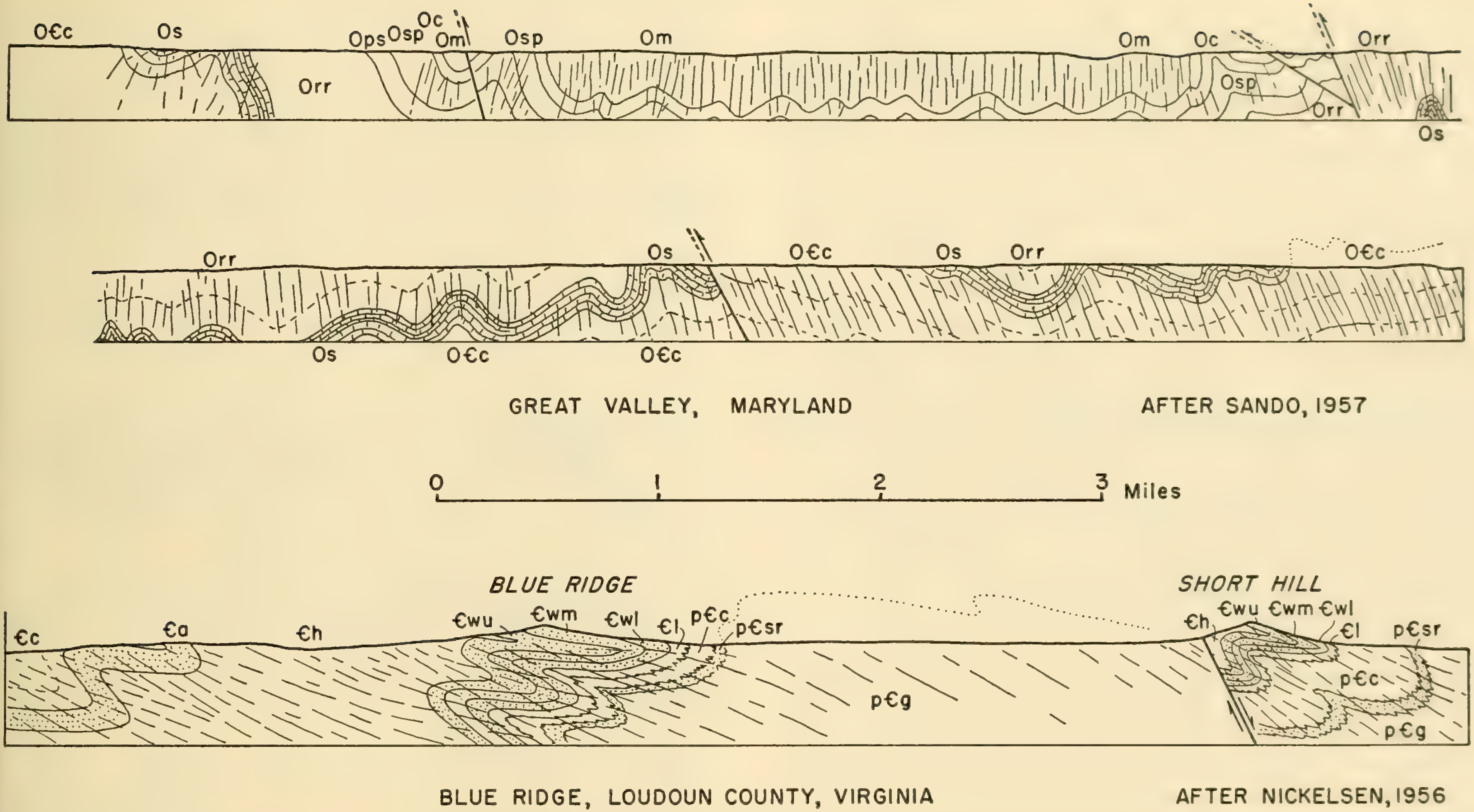


Fig. 8.24. The Great Valley (Shenandoah) and Blue Ridge in Maryland and northern Virginia. For location see Fig. 8.22. Om, Martinsburg shale; Oc, Chambersburg limestone; Osp, St. Paul group (limestone); Ops, Pinesburg Station dolomite; Orr, Rockdale Run formation; Os, Stonehenge limestone; O€s, Conococheague formation; €c, Tomstown formation; €a, Antietam quartzite; €h, Harpers formation; €wu, €wm, and €wl, Weverton quartzite; €l, Loudon formation; p€c, Catoctin metabasalt; p€sr, Swift Run phyllite; p€g, gneissic basement.

chiefly by the Carolina slate belt, which extends from Virginia through the Carolinas into Georgia (see Fig. 7.1). Its rocks are slates, graywackes, pyroclastics, and lavas, which are only moderately folded or metamorphosed except near some granitic body. Still farther southwest in southwestern Georgia and southeastern Alabama is the smaller Pine Mountain belt of quartzite, marble, and schist. The age of the rocks of both the Carolina slate belt and Pine Mountain belt is unknown, but recent workers are inclined to think they may be early Paleozoic and

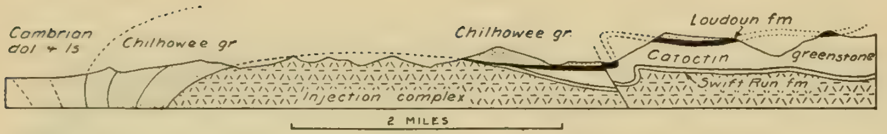


Fig. 8.25. Blue Ridge near Elkton, Va. After P. B. King, 1950b.

somewhat metamorphosed during the later Taconian or Acadian orogenies.

Batholiths

A number of plutons, most of batholithic proportions, occur in the Piedmont province. Their distribution is shown on the *Tectonic Map of the United States,* on the *Geologic Map of the United States,* and on the *Geologic Map of North America.* Major differences in distribution appear on the three maps; the later one shows a far less extent of the plutons in South Carolina and Georgia than the earlier one. According to Keith (1923) most of the plutons are granite and are little deformed or non-deformed. According to Jonas (1932) the Petersburg granite of Virginia is not deformed; it cuts across older structures without disturbing them and enters the rock by replacing those already there.

The plutons are known today, however, to be both concordant and discordant. The former are foliated, and in the older reports are considered early Precambrian. The more or less discordant plutons are the massive ones, and according to the older reports (Keith, 1923, and others) are of late Paleozoic age and associated in time with the folding and thrusting of the Valley and Ridge province. The separation into two vastly different time groups is now held to be unwarranted for two reasons: (1) A similar complex is well-worked out in New England (Chapter 11), and on the basis of fossils and stratigraphic succession the intrusions range in age from Late Ordovician to Carboniferous; (2) isotope age determinations now date the intrusions as Paleozoic. It seems probable that the metamorphism of the Blue Ridge and crystalline Piedmont developed progressively during Paleozoic time as a result of orogeny, possibly several phases of orogeny. The silimanite schist and gneiss zone of the inner Piedmont evolved as a result of the invasion of the vast granitic plutons.

Structure of the Piedmont

From within the central metamorphic and plutonic belt northwestward to the Great Valley nearly all the faults, folds, and cleavage are steeply inclined but have a northwestward asymmetry; i.e., the fault planes, fold axial planes, and cleavage planes dip to the southeast. Toward the Coastal Plain a tendency is noted for the opposite asymmetry. The northwest asymmetry of the inner zone (Fig. 7.2) is more one of foliation than major displacement along a few discrete faults, with relatively slight movement along an infinite succession of foliation planes (Bloomer, 1950).

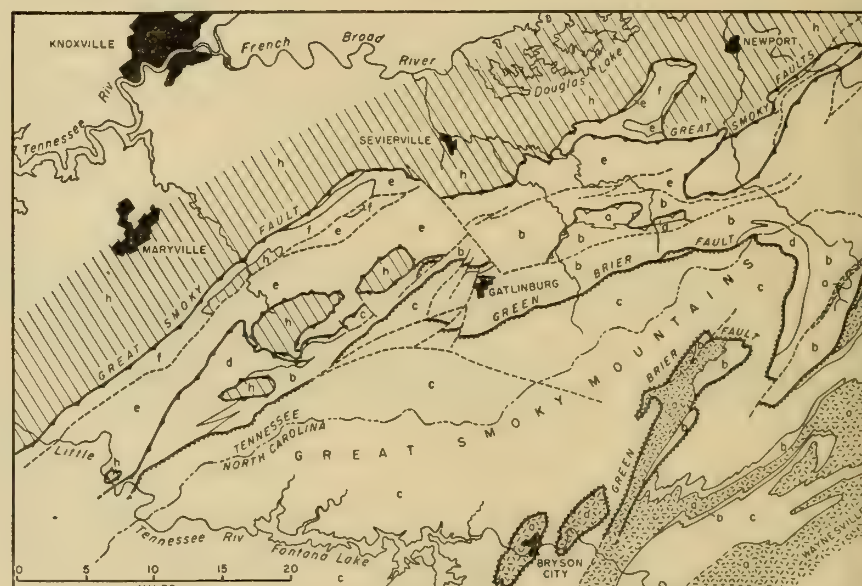


Fig. 8.26. Geologic map of Great Smoky Mountains and vicinity. After King *et al.*, 1958. A, Early Precambrian granitic and gneissic rocks; b,c,d,e, groups of the Ocoee series (later Precambrian); P. Chilhowee group (Cambrian and Precambrian(?)); h, Mississippian, Ordovician, and Cambrian rocks.

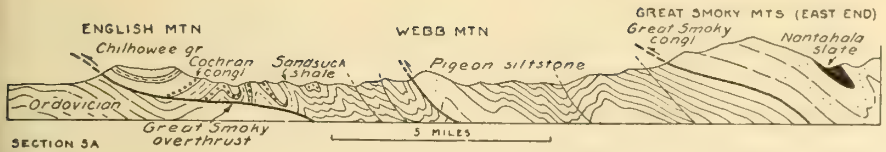


Fig. 8.27. Northeast part of Great Smoky Mountains and adjacent foothills on north. After P. B. King, 1950a. The Great Smoky conglomerate, the Nantahala slate, the Pigeon siltstone, and the Sandsuck shale, are part of the Ocoee series (Late Precambrian) which forms most of the Great Smokies. The Cochran conglomerate is basal Cambrian. For location see Fig. 8.22.

Infolded Belts of Metasedimentary Rocks

Besides the gneisses, the metamorphic and plutonic belt contains other metamorphic rocks that are clearly of sedimentary origin. These characteristically form narrow belts or bands of considerable linear extent. The principal belts of metasedimentary rocks are:

1. The Arvonia slate belt, near the James River, and the Quantico slate belt, near the Potomac River, in Virginia. These are synclines of fossiliferous Ordovician rocks, lying uncomformably on older schists and granites.
2. A belt of quartzite, schist, and marble in North and South Carolina, which has been mapped by Keith (1931) in the Kings Mountain area. Further details have been given by Kesler (1944), whose interpretations differ from those of Keith.
3. The Brevard schist belt [Figs. 7.1 and 8.30], which is by far the longest, and extends from central North Carolina through South Carolina, Georgia, and Alabama to the Gulf Coastal Plain. Jonas (1932) states that similar rocks continue northeastward from central North Carolina into southern Virginia. The rocks of the Brevard belt consist of contorted dark slates and schists, with lenses of limestones, apparently of a somewhat lower grade of metamorphism than the rocks which flank them on either side.
4. The Murphy marble belt of western North Carolina and Northwest Georgia (Fig. 8.22), has many features similar to the others just described, but differs in that it is not flanked by crystalline rocks, but by altered sedimentary rocks of the Ocoee series.

No fossils have been found in the belts south of Virginia and the age of the rocks which compose them is unknown. They have been variously assigned to the Paleozoic and the Precambrian (King, 1950a).

Carolina Slate Belt

In the southeast part of the Piedmont province, highly metamorphosed rocks give place to less metamorphosed sedimentary and volcanic rocks which make up the Carolina slate belt (Fig. 7.1). Granite intrusions are present, but they appear to be small and widely scattered and also crosscutting rather than concordant. The most extensive rock unit is the "volcanic series." It is composed of flows, breccias, and bedded tuffs of volcanic origin with some interbedded slates and sandstones. To the southwest in southwestern Georgia and northeastern Alabama is the shorter and narrower Pine Mountain belt. It is composed of quartzite, marble, and schist clearly of sedimentary origin and intruded by a gneissic granite. The beds are broadly rather than steeply folded. The age of both the rocks of the Carolina slate belt and the Pine Mountain belt is uncertain; they have been assigned to both the Precambrian and Paleozoic.

Paleozoic of Florida

Within the area embracing northern Florida and adjacent parts of southern Alabama and Georgia, recent drilling has shown that the Mesozoic rocks are underlain by volcanic rocks and by sedimentary rocks of Paleozoic age (Applin, 1949).

In the Ocala uplift, pre-Mesozoic rocks are reached in places at depths of less than 4000 feet, but elsewhere they may lie as much as 10,000 feet below the surface. Penetration of the pre-Mesozoic rocks has not been sufficient to establish a sequence; in other words, different rock types have been found in different wells, but have not been found in superposition.

The sedimentary rocks are mainly sandstones and shales. Some of the sandstones contain worm tubes of *Scolithus* type, not unlike those found in the older Paleozoic rocks of the Appalachians; others contain large quantities of detrital mica. The shales are gray, black, and even red. Graptolites have been found in places, as well as various other fossils. The only Paleozoic systems whose existence has been definitely proved paleontologically are the Ordovician and Silurian, although others might be present. The volcanics may be related to the "volcanic series" of the southeast part of the Piedmont area, but like this series, their age has not been established.

Well cores show that these rocks are little deformed. Metamorphic effects, such as cleavage and recrystallization, are lacking. Bedding dips

Fig. 8.28. Serpentine belt of the Appalachians. By H. H. Hess, Princeton University; and published with his permission. Circles represent known bodies of serpentine.

at low angles; in some places the strata are flat, and the maximum inclination is 25° to 30°. Drilling is too widely spaced to permit determination of more than the gross structural pattern. As the rocks have been encountered over an extensive area, even these low dips would be sufficient to account for a sedimentary and volcanic sequence of considerable thickness.

These discoveries are of great interest, as they show that southeast of the Appalachian system there is a foreland or shelf of little deformed rocks, just as there is northwest of it.

Ultrabasic Intrusives

Hess (1937a) has charted the serpentinized ultramafic intrusives of the Appalachians and finds they form a narrow belt lengthwise of the Piedmont crystalline province through New England to Quebec City, thence through the Taconic and Acadian belt of Quebec to the Gaspé Peninsula, and again in a belt through Newfoundland. See the map of Fig. 8.28. In his work in the Greater and Lesser Antilles, he has concluded on the basis of considerable evidence (see Chapter 42) that the serpentines occur in the arcuate, highly deformed, orogenic belt, and as a conclusion, that in certain ancient orogenic belts, now obscured by metamorphism and blanketing deposits, they can be taken to indicate the position of the zone of maximum orogeny. The serpentinites are chiefly associated with the Taconian orogenic belt in New England and the Maritime Provinces, and they are strong evidence, it seems to Hess, that the core of the Taconian orogeny stretched through the crystalline Piedmont of the southern and central Appalachians.

Besides the granite plutons, the metamorphic and plutonic belt contains a group of intrusives of ultrabasic composition—peridotites, dunites, pyroxenites, and others, now in part altered to serpentine. Unlike the granites, they mostly occupy small areas, but in many places they form well-defined zones, indicating that they were intruded under the influence of some sort of tectonic control. The most prominent zone lies toward the northwest edge of the metamorphic and plutonic belt, in the southeast part of the Blue Ridge province of western North Carolina; it continues northeastward into Virginia, and southwestward into Georgia.

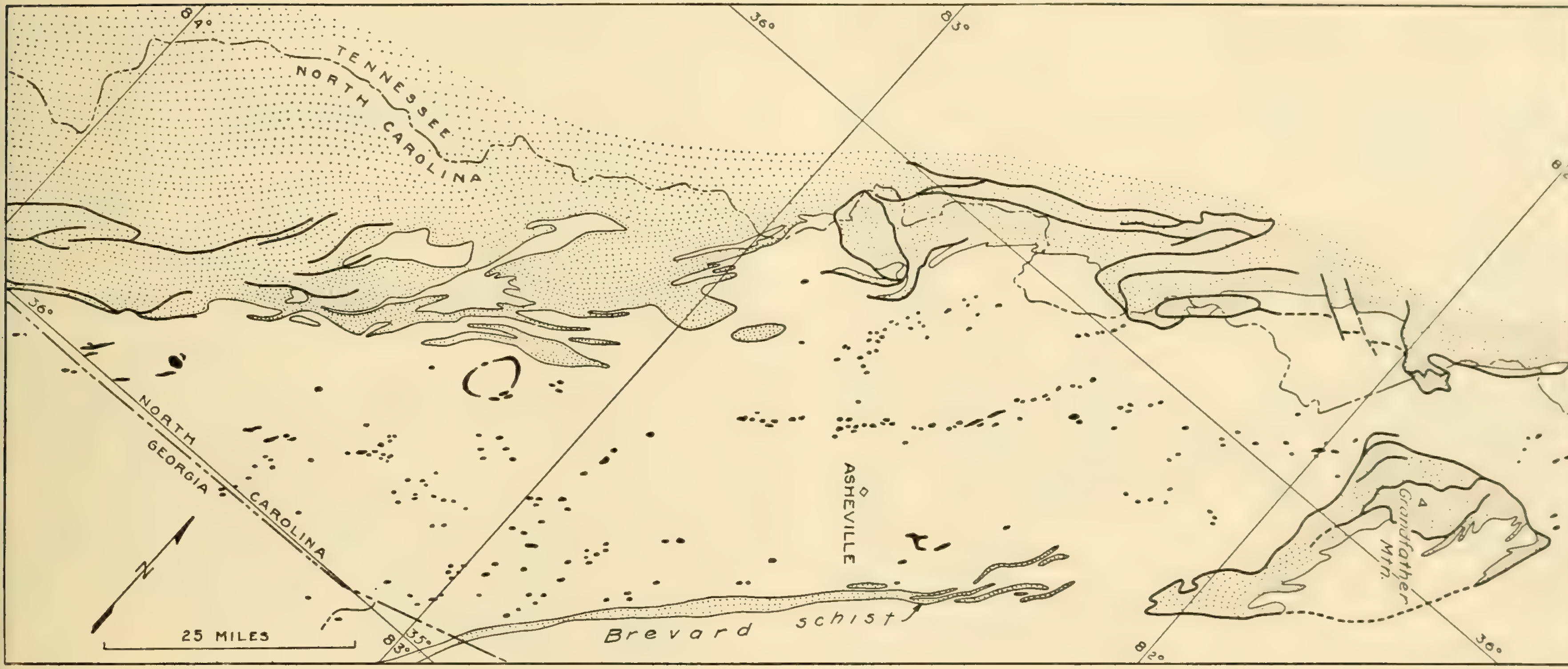


Fig. 8.29. Map of part of the Blue Ridge and Piedmont provinces of western North Carolina, showing the distribution of ultrabasic igneous rocks. After P. B. King, 1950a. Stippled areas are those of sedimentary rocks, mainly Paleozoic, but including the Ocoee series, probably Precambrian, and Brevard schist of unknown age. Blank areas east of, and within, stippled area are those of gneiss and schist with bodies of intrusive granite. Small black areas are ultrabasic igneous rocks. Heavy lines are faults. After P. B. King, 1950a.

Other less well-defined groups of intrusives occur toward the southwest edge of the metamorphic and plutonic belt, as near the inner margin of the Coastal Plain in Georgia. The age of the ultrabasic intrusives in the southern Appalachians is unknown. Pratt and Lewis, on very tenuous evidence, conclude that they are of older Paleozoic age (King, 1950b). See Fig. 8.29.

Crystallines of Maryland and Southern Pennsylvania

The Piedmont of Maryland and southern Pennsylvania merits special attention because of the considerable detailed work done there by Ernst Cloos, students, and colleagues. Cloos (1953) has divided the region into twelve divisions or zones, the first being the Coastal Plain. See map, Fig. 8.30. The second division is the belt of most intense metamorphism of the Piedmont province (Wasserburg *et al.*, 1957) and contains a number of gneiss domes. Six of these are in the vicinity of Baltimore and their cores are made up of gneiss and migmatite (Baltimore gneiss) which are mantled by the metasediments of the Glenarm series (Tilton *et al.*, 1958). The lowest formation of the Glenarm is the Setters quartzite, the next above the Cockeysville marble, and the last the Wissahickon

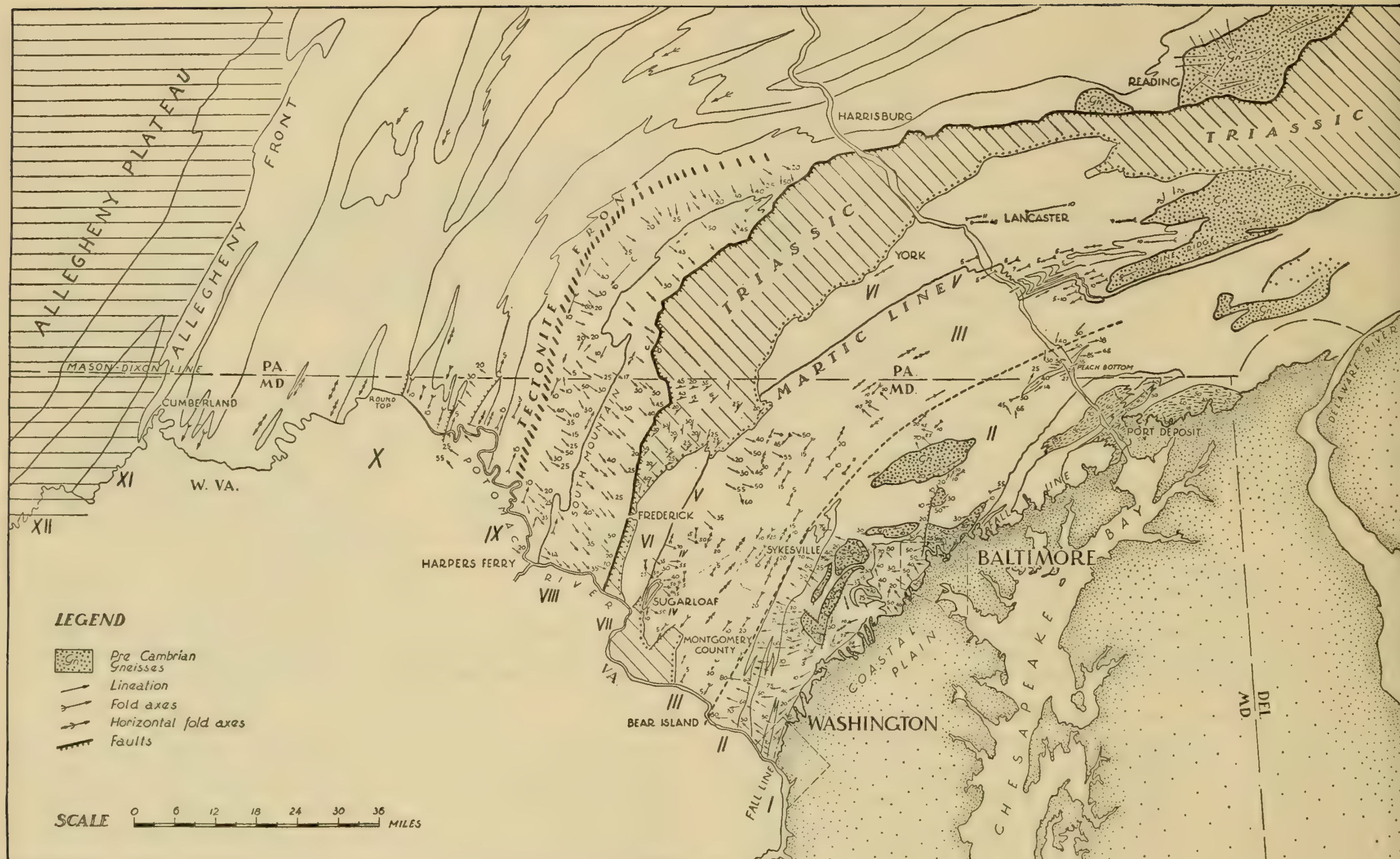


Fig. 8.30. Tectonic map of Maryland and southern Pennsylvania. Reproduced from Cloos, 1953.

schist. Granitic stocks and pegmatite dikes cut the domes and metasedimentary mantle.

Foliation in the Baltimore gneiss parallels the contact with the mantle and arches over the domes in asymmetrical form with the steep flanks to the southeast. Lineation appears like raindrops running off an umbrella (Cloos, 1953).

The Baltimore gneiss has been considered Precambrian in age and possibly as old as any rock in the Piedmont. The Glenarm sediments are thought to have been deposited in late Precambrian or early Paleozoic time unconformably upon the gneiss. It is clear, however, that the same degree of metamorphism pervades the overlying Glenarm rocks as the Baltimore gneiss, and since the foliation of one parallels the other, it has been assumed that metamorphism and doming of the mantle has obliterated the original basement structures and produced a new concordant foliation.

If the unconformity existed, two periods of tectonism are implied, one prior to Glenarm sedimentation and another following it. If the unconformity did not exist, a single period of deformation, metamorphism, and injection can explain observed relationships. All previous investigators of the domes favor existence of the unconformity, but conclusive proof is lacking (Tilton *et al.*, 1958).

The age of the post-Glenarm tectonism is generally considered Taconian or Acadian. Evidence bearing on this conclusion will be presented later. The age of the Precambrian tectonism will also be taken up later.

The third division of the Piedmont shown on Fig. 8.26 consists mostly of the Glenarm series with generally horizontal fold axes. Foliation is vertical on the east border and is inclined to the southeast on the west border. The fold axial planes dip to the southeast also. Metamorphism lessens toward the west, and mica schists become phyllites; amphibolites become epidote- and chlorite-rich greenschists.

The rocks of this zone have not yet been correlated with the fossil-bearing early Paleozoic strata west of the Martic line. It is possible, according to Cloos, that the Cambro–Ordovician limestones of the westerly zone (6) are facies of the once sandy rocks of the Glenarm.

Zone four encompasses the Sugarloaf structure which, as shown by the closure of the bedding, is an anticlinal dome. The western limb is overturned. Cleavage confirms the domal structure. The rocks are in the chlorite and greenschist facies of zone three. The local phyllites are correlated with the Cambrian Harpers phyllite, and the quartzites which are below the Harpers are most likely the Lower Cambrian Weverton quartzite (Scotford, 1951).

The Martic line is called division five. It was first recognized as an overthrust in which the then presumed older Wissahickon schist was thrust westward over the presumed younger Paleozoic strata, and all rocks southeast of the "fault" were regarded as Precambrian and northwest of it as Paleozoic. Careful work has shown that the line is not a discrete plane of major displacement, but that in most places complicated conditions pertain (Cloos and Heitanen, 1941). It was also presumed that the Martic "thrust" is a boundary between highly metamorphosed schists and little metamorphosed Paleozoic strata. Cloos and Heitanen have demonstrated that metamorphism is not restricted to the Wissahickon schist but that all rocks including the Cambrian Antietam schist, Vintage dolomite, and Ordovician Conestoga limestone show the same intensity of metamorphism. At one place the sequence is repeated five times, where the Conestoga is capped by the Wissahickon schist, which in several ways is similar to the Antietam. At another place the Antietam schist almost meets a spur of Wissahickon.

Along the Martic line the fold axes are horizontal or plunge predominantly to the southwest. All folds are overturned southward. Flow cleavage is an axial plane cleavage and dips to the north. Bedding is intensely crumpled and at many localities is entirely obscured by later cleavage. Since all members of the sequence are thin and underlie large areas, it can safely be assumed that bedding is roughly parallel to the boundary planes and thus largely conformable in all members of the sequence (Cloos and Heitanen, 1941).

Zone six consists mainly of Cambro–Ordovician limestones which are strongly cleaved and overturned to the west. The zone is covered in large part with the Triassic deposits (division seven).

Zone eight is the Blue Ridge belt previously described, and cleavage and lineation extend northwestward to the position labeled "tectonite

front." From this line westward the sedimentary rocks are non-tectonites. The other zones have been described in previous parts of this chapter.

Age Determinations by Radioactivity

The first isotope age determinations on the minerals of the crystalline Piedmont were published in 1941 (Goodman and Evans), and since then methods and calculations have been refined, new methods developed, and a fair number of presumably reliable dates have been determined.

Two groups of ages are now fairly well established, namely, one ranging from 1000 to 1100 m.y. and one ranging from 250 to 390 m.y. References to all significant dates may be found in recent publications by Tilton *et al.* (1958), Hurley *et al.* (1958, 1959), and Carr and Kulp (1957). An abstract by Kulp *et al.* (1957) is significant for ages in the southern Piedmont.

The older ages (1000–1100 m.y.) come principally from zircon subjected to $\frac{U238}{Pb206}$, $\frac{U235}{Pb207}$, $\frac{Pb207}{Pb206}$, and $\frac{Th232}{Pb208}$ analyses.

Three of the mantled domes in the Baltimore area (zircon from the Baltimore gneiss), two gneisses from Bear Mountain, New York, a gneiss from Shenandoah National Park, Virginia, and two gneisses from Hibernia, New York, were sampled and the zircons run. The results range from 1030 to 1170. Rubidium-strontium age measurements were also made on microcline from the three Baltimore gneiss domes and a value is fixed for one at 1200 plus 100 or minus 200 m.y. and for another at about 1040 m.y. It is concluded by Tilton *et al.* (1958) that the zircon and microcline ages record a 1000–1100-m.y. crystallization in the Piedmont.

Now, from the same specimens of Baltimore gneiss from which the zircon and microcline ages were obtained, biotite by $\frac{K40}{A40}$ and $\frac{Rb87}{Sr87}$ analyses gave ages of 305–339 m.y. For the older and younger dates of the same rock two interpretations can be thought of:

(1) The gneiss was crystallized or recrystallized 1000–1100 m.y. ago (2) The gneiss was originally a clastic sediment metamorphosed 300–350 m.y. ago, and the zircon and microcline were relict detrital grains eroded from a terrain 1000–1100 m.y. old. The first interpretation is favored, chiefly because of the non-clastic character of the microcline grains. Their irregular shapes, with delicate projections and interlocking contacts with other minerals, were clearly formed during crystallization of the gneiss. Possible detrital origin for the zircon cannot be excluded, although if this were the case a greater age than that of the microcline might be expected. It is concluded that the microcline and the zircon probably record a 1000–1100 m.y. crystallization in the Baltimore gneiss, while biotite records a second crystallization 300–350 m.y. ago. It should be noted that these conclusions allow either a sedimentary or igneous origin for the gneiss (Tilton *et al.*, 1958).

Kulp *et al.* (1957) report a granite from eastern Georgia about 250 m.y. old. They also give "apparent ages" of 320–370 m.y. for the "metamorphic series" in western Virginia and North Carolina as well as the pegmatite swarms in the Spruce Pine and Bryson City districts of North Carolina.

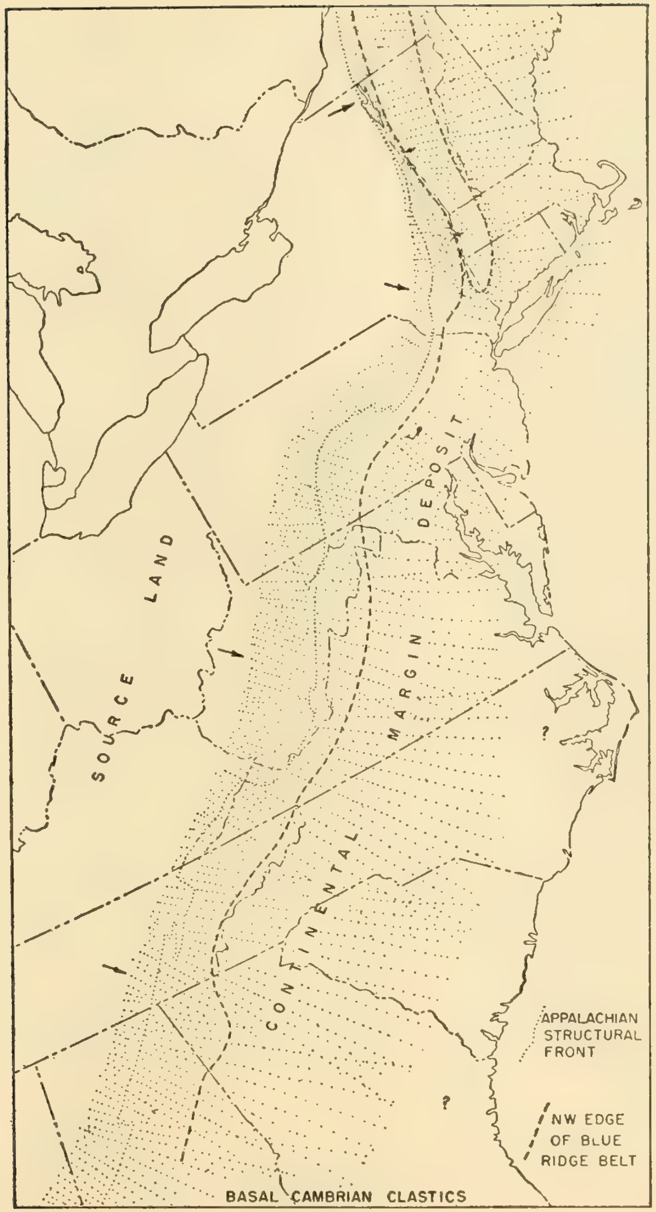
In New England a number of radioactivity age measurements have been made on plutons where the intrusive relations to well-dated fossiliferous strata are visible, and it is concluded that the Devonian period began approximately 400 m.y. ago and ended slightly less than 250 m.y. ago. These data will be presented in Chapter 11 on New England. It appears, therefore, that the recrystallization and plutonism (tectonism) in the Piedmont province ran its course during the Devonian period. This is younger than the Taconian orogeny of New England and the Maritime provinces which, from angular unconformities, is dated as late Ordovician. The Acadian orogeny is generally regarded as having occurred during the upper half of Devonian time, so the dates over 300 m.y. seem too old for it, unless extended by definition.

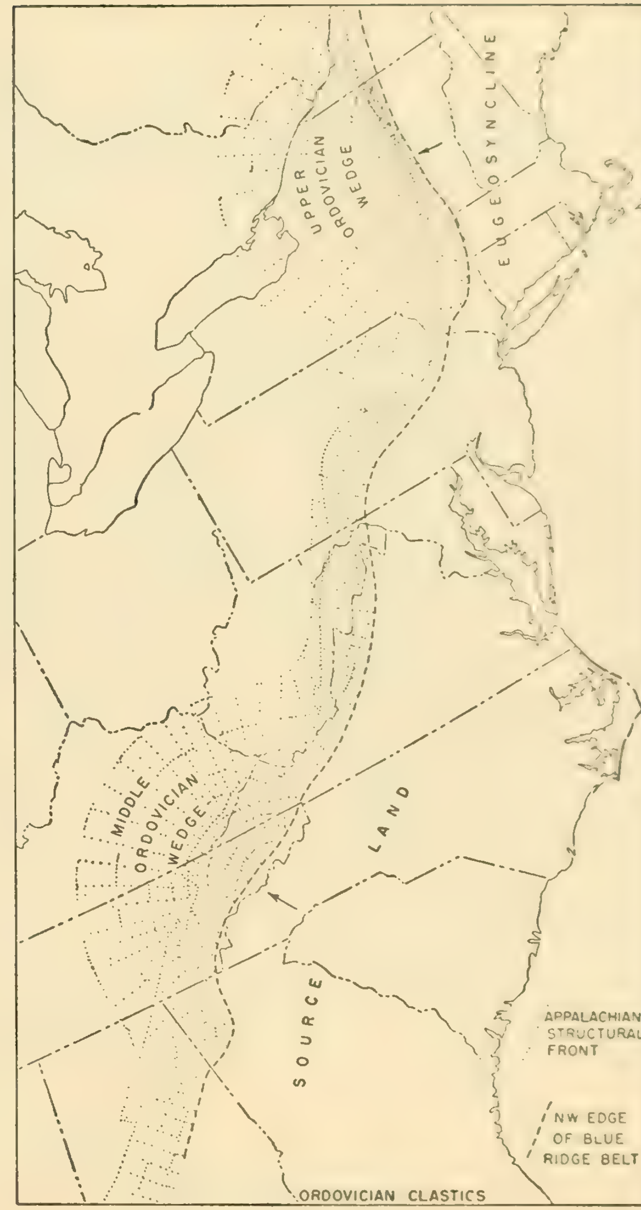
SUMMARY OF OROGENIC HISTORY

The major lines of evidence of orogeny in the Appalachian mountain system come from the sedimentary domains, the structures and structural relations, metamorphism, plutonism, and isotope age determinations. These have all been reviewed, and now may be integrated and the following conclusions reached.

1. An orogeny occurred along the Atlantic margin of the United States south of New York City in which previously existing rocks were

Fig. 8.31. Regimen of Appalachian sedimentation in the early Paleozoic. Partly after P. B. King, 1959.





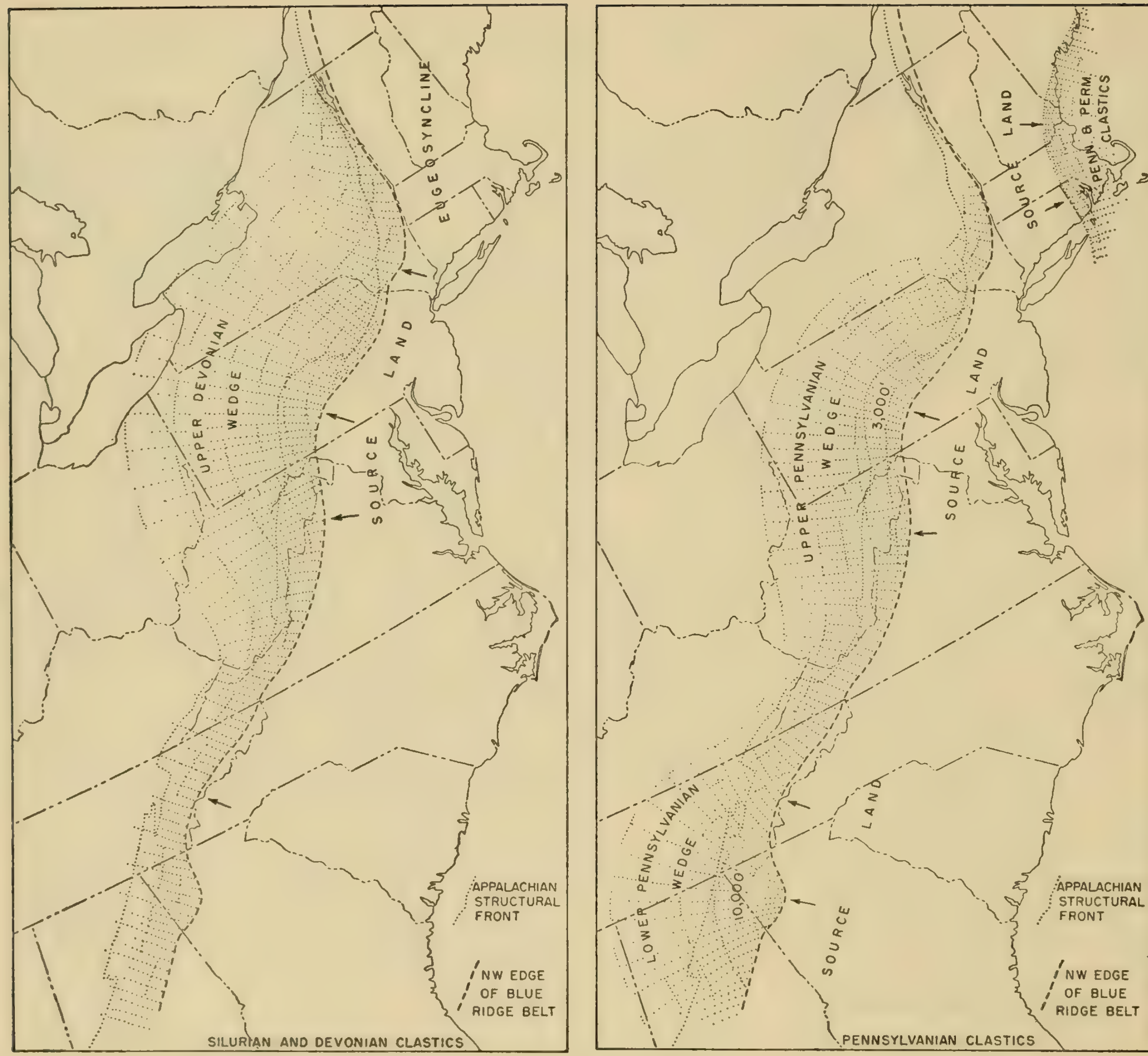


Fig. 8.32. Regimen of Appalachian sedimentation during Middle and Late Paleozoic. Partly after P. B. King, 1959.

recrystallized 1000–1100 m.y. ago. This correlates in time with the Grenville orogeny of Ontario and Quebec.

2. The continental margin was subparallel with the present, but may have been extended by a continental shelf and slope type of deposit in times following the Grenville orogeny, particularly in Late Precambrian and Early Cambrian time. This was the time of accumulation of the Ocoee series and the Chilhowee group.

3. The Atlantic margin of the continent was beset with deformation beginning in the last part of early Ordovician time, and the previous region of sedimentation now was elevated and became the source land of sediments to the west. See Fig. 8.31. A great fan or wedge of clastic sediment was spread northwesterly from the Great Smoky region during the Middle Ordovician and another one in Late Ordovician time in New England. The crustal deformation must have been mostly elevatory at this time because the metamorphic and plutonic activity occurred somewhat later. The New England clastic wedge records part of the Taconian orogeny as defined, but no name has been proposed for the Middle Ordovician uplift.

4. Clastic sedimentation on a large scale shifted during Silurian and Devonian time to New York, Pennsylvania, and West Virginia, and another great fan of sediments was deposited there, also derived from uplifted lands on the east. See Fig. 8.32. The Silurian and Lower Devonian clastics were not very thick, about 5000 feet, but then a flood of sediments reached 10,000 feet in thickness in late Devonian time. Strong compression and plutonic tectonism started in early Devonian time, according to the isotope age measurements, but evidently high mountains were not created until the beginning of the Late Devonian.

Figure 8.33 is an idealized section of the southern Appalachian system and illustrates the central belt of most profound Devonian metamorphism and plutonism. This, when much eroded, became the crystalline Piedmont. The age of the Carolina slate belt sediments is unknown but evidently older than the Devonian tectonism. It may be speculated that they were a collateral eastern deposit of the Middle Devonian clastic wedge on the west of a medial uplift, but their age must be known first before they can be correctly fitted into the picture.

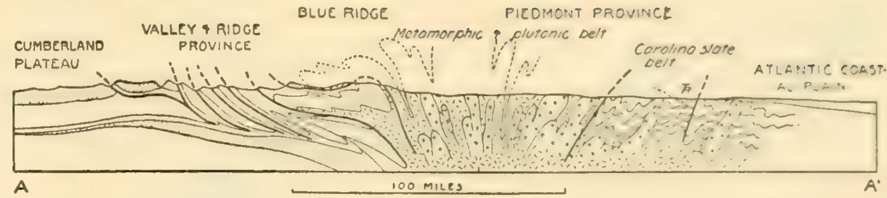


Fig. 8.33. Idealized cross section of the southern Appalachian Mountains system. After P. B. King, 1950a.

5. Uplift of the orogenic belt was general along its entire length during the Mississippian and sediments were carried westward and added to the miogeosyncline. However, in Early Pennsylvanian time uplift was particularly great in the southern Piedmont and another thick wedge accumulated on the west. Later, sedimentation shifted to the West Virginia and New York and considerable clastic material of continental environment accumulated during Late Pennsylvanian and Permian time.

6. The eastern part of the miogeosyncline including the thickest parts of the clastic wedges and the eastern part of the great Cambro–Ordovician carbonate sequence was then compressed and cast into folds and thrusts as exemplified in the Valley and Ridge province of Fig. 8.33. The deformation is generally referred to as the Appalachian orogeny. It may have started in Mid- or Late Pennsylvanian time in the south but farther north Valley and Ridge deformation could not have occurred until the close of Pennsylvanian time, and it may not have happened until near the close of the Permian.

9.

EASTERN TRIASSIC BASINS

DISTRIBUTION OF BASINS

A series of long, narrow basins of Triassic deposits occurs along the eastern margin of the continent. It will be seen by reference to the *Geologic Map of the United States* or *Geologic Map of North America* that the basins start at the north boundary of South Carolina in the Piedmont crystalline province and extend through North Carolina, Virginia, Maryland, Pennsylvania, and New Jersey to the lower Hudson River Valley in New York. The basin in Pennsylvania and New Jersey is the largest of any in the United States, and for a distance between the Carlisle prong and Reading prong of the Blue Ridge element, it borders on the Ridge and Valley province.

The Connecticut River Valley in Connecticut and Massachusetts is the site of a large Triassic basin, and under the Bay of Fundy and along its east shore in Nova Scotia another such basin exists. See Plate 9.

The Triassic areas are generally sites of lowlands because the basin beds have yielded to erosion more than the adjacent crystallines. The Triassic lowlands is the physiographic name generally given to the Pennsylvania–New Jersey basin. The lowlands are marked, however, by ridges of trap rock that stand rather prominently above the lowland plain.

NATURE OF TRIASSIC ROCKS

General Character

The Triassic sedimentary rocks of the eastern basins are chiefly clastic and dominantly red. Fanglomerates, conglomerates, sandstones, arkoses, siltstones, shales, and argillites are the common sedimentary types. Much basic magma has invaded the sediments and now exists as thick sills and long dikes of diabase. Basalt flows from the same magma are also intercalated in the shales and sandstones. The intrusive rocks have commonly altered the red sediments to blue or gray along the contacts in zones 50 to several hundred feet thick.

New Jersey–Pennsylvania–Maryland–Virgina Basin

Newark Group. The sediments of the New Jersey–Pennsylvania–Maryland–Virginia basin are known as the Newark group. The basin has a maximum width of 30 miles and is over 300 miles long. Part of it is shown in Fig. 9.1. The Newark group has been classified in three formations, the Stockton, Lockatong, and Brunswick, the last-named being the youngest. These subdivisions are clearly separable along the Delaware River and northeastward in New Jersey, where they were first established and named.

The Stockton formation in general comprises arkosic sandstone with some red-brown sandstone and red shale, in irregular succession and presenting many local variations in stratigraphy. It lies unconformably on Paleozoic and pre-Paleozoic crystalline rocks. The sandstones are in places cross-bedded, and the finer-grained rocks exhibit ripple marks, mud cracks, and raindrop impressions, which indicate shallow-water conditions

during deposition. The arkose, a sandstone containing more or less feldspar or kaolin derived from granite or gneiss, indicates proximity at the time of deposition to a shore of Precambrian crystalline rocks.

The Lockatong formation consists chiefly of dark-colored fine-grained hard and compact argillaceous rocks. Some beds are massive, and others are flaggy. They show mud cracks and other evidences of shallow-water deposition, but their materials are clay and very fine sand, some of the beds also contain carbonaceous material.

The Brunswick formation, in its typical development, consists mainly of a great thickness of soft red shale with local and thin layers of sandstone. Northward and westward the sandstone increases in amount and coarseness. It overlaps irregularly older Traissic formations and Paleozoic and pre-Paleozoic formations.

The three formations are not sharply separated by abrupt changes of material, but usually merge into one another through beds of passage which appear to vary somewhat in thickness and possibly also in stratigraphic position in different areas.

The thickness of the Stockton is estimated to range from 1000 to 3000 feet, the Lockatong from 1500 to 3000 feet, and the Brunswick from 12,000 to 16,000 feet. The total thickness of the Newark group as generally mentioned is about 20,000 feet, but figures up to 35,000 feet have been proposed. This great amount is computed by the dip angle and the distance across dip of the homoclinal beds, but several writers have suggested the possibility of duplication of certain beds by faulting, and hence that the figure may be excessive. Stose and Stose (1944) suggest that the beds overlapped from east to west in somewhat the manner shown in Fig. 9.2 and that therefore the combined thickness of all the beds will not be found in any one place. It cannot be doubted, however, that the long, narrow troughs containing the Triassic sediments are very deep, undoubtedly over 10,000 feet, and probably 20,000 in places.

The age of the Newark group is probably Upper Triassic, but the highest beds may be lowermost Jurassic. According to Bascom and Stose (1938),

> A comparison of fossil plants, crustaceans, and vertebrates of the Newark with simliar forms of the Jura and Trias of Europe establishes a correspondence within general limits, but a correlation of exact horizons is not practicable.

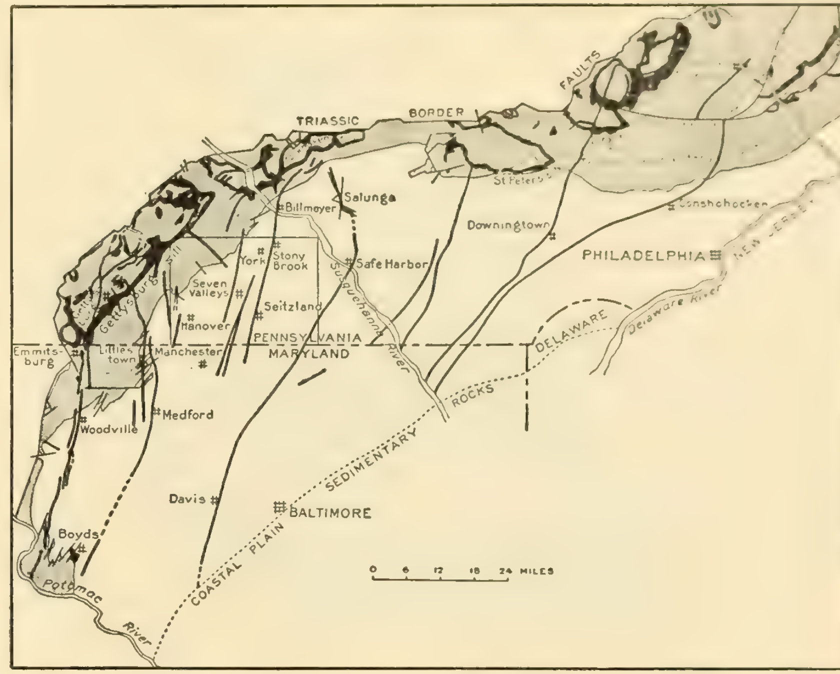


Fig. 9.1. Triassic basin in western New Jersey, Pennsylvania, and Maryland. Stippled area, Triassic sedimentary rocks; solid black areas and heavy black lines, Triassic diabase sills and dikes; light black lines, faults. Reproduced from Stose and Stose, 1944.

The Newark strata did not share in the folding that occurred at the end of Carboniferous time and therefore must be of later date; they are, however, clearly older than the lowest Cretaceous formations, which overlap them unconformably. They are thus separated from earlier and later deposits by intervals of upheaval and erosion of unknown duration, but their position in geologic history cannot be determined more closely than by the general correlation of fossils above indicated.

Igneous Rocks. The map of Fig. 9.1 shows the distribution of outcropping sills, lava flows, and dikes in the Newark group and in adjacent rocks of the Piedmont. The sills and flows are confined to the Triassic basin, but some of the dikes cross out into the older rocks of the Piedmont and persist for many miles. The Conshohocken and Downington dikes

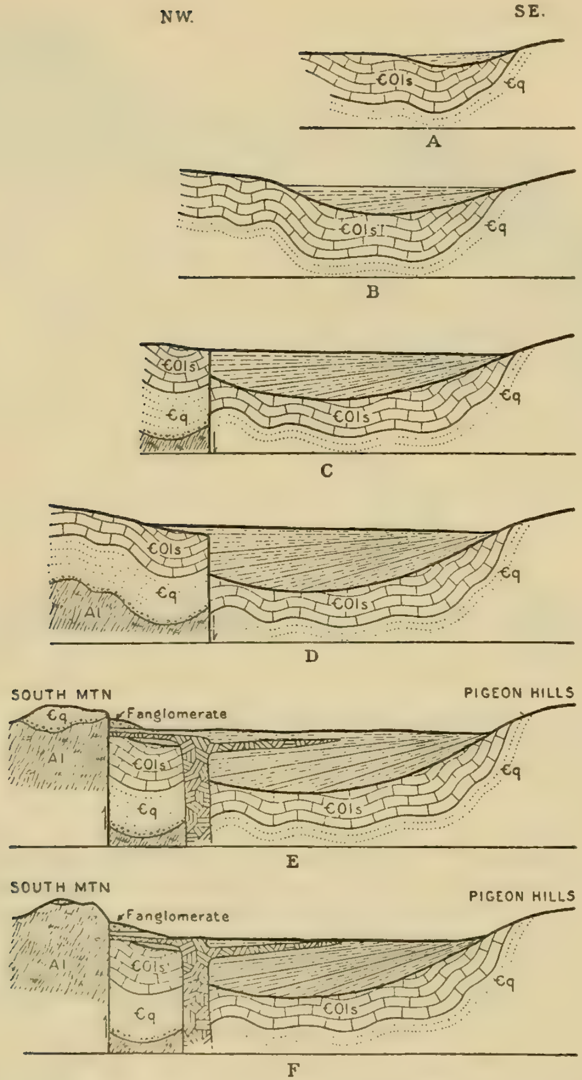


Fig. 9.2. Origin of the Newark Triassic basin. Reproduced from Stose and Stose, 1944. The sediments are postulated to have first been derived almost entirely from the east. After intrusion of the diabase dikes and sills and renewed faulting, much fanglomerate was washed in from the west.

are 60 to 70 miles long, and the Safe Harbor dike extends an equal length before it is covered by the Cretaceous of the Coastal Plain.

The largest sill in the southern part of the Newark basin is the Gettysburg, which is 1800 feet thick. Farther northeast in New Jersey four great sheets of trap rock occur and form the Watchung Mountains which are more prominent that the ridges of the great Gettysburg sill. The lowest of the four sheets is intrusive and in places reaches a thickness of 1000 feet. It forms the Palisades of the Hudson. See section 31 of Fig. 8.21. Above the Palisades sill and separated from it and each other by several hundred feet of intervening Triassic shales are three extensive (buried) basalt flows which, from bottom to top, are 650, 850, and 350 feet thick.

The dikes are believed to follow tension cracks which in places become faults and offset the Triassic beds. According to Stose and Stose (1944) the dikes and the normal faults that the dikes follow represent major lines of Triassic fractures. They cut across older structural lines, which are nearly at right angles to them. Many of the diabase dikes originate in or join the diabase sills which are most abundant along the northwestern part of the basin. See map, Fig. 9.1.

The diabase sills, with which many of the dikes connect at their northwestern ends, coalesce to form extensive intrusive bodies in the northwestern part of the Triassic area of Pennsylvania. The larger sills are the Haycock, Ziegler, Saint Peters, Yorkhaven, and Gettysburg. They parallel the strike of the sedimentary rocks for long distances, and then the intrusive body cuts across the strike at right angles. Most of these crosscutting bodies extend to the northwestern edge of the Triassic basin where they terminate against the faults that form the boundary of the basin. Each of these intrusive bodies, therefore, has the form of a great tilted trough bounded on the southeast side by the west-dipping sills and at the ends by the crosscutting bodies and open at the west.

The fissures through which the diabase entered the Triassic rocks are believed to lie near the northwest edge of the basin where the greatest amount of progressive sinking and faulting occurred during Triassic deposition. The rising magma broke through the Triassic beds near the vents in the form of crosscutting bodies, and injected the beds to the southeast in the form of sills. The magma extended still farther southeastward as dikes that followed vertical fractures in the Triassic sedimentary rocks and continued into the older underlying rocks southeast of the limits of the basin of Triassic sedimentation. Some of these dikes in the area southeast of the Triassic outcrops may have been feeders of large diabase bodies in Triassic sedimentary rocks that are now re-

moved by erosion, but the evidence is not available to support such a view (Stose and Stose, 1944).

Border Conglomerate. Along the northwest border of the Triassic basin occur deposits of fanglomerate, generally called conglomerate and breccia. They make up the "Border conglomerates." In width of exposure they range from less than half a mile to about 8 miles and lie in discontinuous patches along the Precambrian and lower Paleozoic rocks of the northwestern border. The largest area is south of Reading, Pennsylvania, which extends across the Gettysburg (Brunswick) formation to the New Oxford (Stockton). Most of the gravel fragments were derived from Lower Paleozoic limestones, dolomites, sandstones, and quartzites, but some came from beds as high as the Devonian, and some are Precambrian rocks. In one place Triassic basalt forms boulders and cobbles in the fanglomerate (Carlston, 1946).

The Border conglomerate is for the most part of Brunswick age, and as depicted in certain cross sections is the top and youngest layer of the Triassic group. It seems to lie unconformably across the older Triassic beds in places, and in others rests directly on the pre-Triassic. On the other hand, the conglomerate beds pass into sandstones and shales and are undoubtedly mostly a northwestward marginal facies of the Brunswick. Even Border conglomerate wedges have been observed in the Stockton and Lockatong, and although the conglomerate is chiefly of Brunswick age, local bodies of it may be of any age within the Newark group (McLaughlin, 1931, 1958).

Although the Border conglomerates clearly betray a northwest origin, most of the material washed into the Triassic basin is thought to come from the southeast. The reason, according to Stose and Bascom (1929) lies in the composition of the basin beds. The "poorly assorted arkosic grits, containing feldspar and mica derived from disintegrating granitic rocks" were exposed, they believe, only in the land southeast of the basin. Except for a stretch of about 75 miles in southeastern Pennsylvania to which Stose and Bascom refer specifically, the Triassic basins in the Piedmont are bordered on both sides by crystalline rocks that could have supplied feldspar and mica, but the Paleozoic pebbles in the border conglomerate indicate that little Precambrian was exposed on the northwest at the time of Triassic deposition in the southeastern Pennsylvania area.

Deep River Basin

The Deep River basin is in North Carolina and is generally regarded as made up of the Cumnock basin on the southwest and the Durham basin on the northeast. The southwestern basin is noted for its Triassic coal. The deposits in these basins are much like those of the Newark basin with an abundance of gray arkosic beds lensing into red sandstones and shales and gray to buff sandstones. Locally thin carbonaceous shale beds occur. Conglomerates, fanglomerates, and in places landslide breccias mark the border zones, but here, unlike in the Newark basin, both borders are marked by the coarse deposits. Thin conglomerates with an abundance of quartz pebbles occur also in the central areas (Prouty, 1931).

The torrential fanglomerates are more voluminous along the eastern margin of the basin than the west, which shows that the eastern margin was the steeper and that an area of land existed there as well as on the west.

Connecticut Valley Basin

The Triassic sedimentary rocks of the Connecticut Valley are all clastic and, if anything, coarser than those in New Jersey, Pennsylvania, and Maryland. Red colors dominate, and they are also interlayered with traprock sheets. The basin is bordered in part on both sides by faults, and is thus a graben; but the eastern fault is by far the greatest and is known as the Great Fault. All beds dip generally eastward into it, as the beds dip generally westward into the border fault of the Newark basin. See map of Fig. 9.3. The Great Fault has a throw estimated variously between 17,000 and 35,000 feet, but the basin beds and floor have not been regarded in the same way as Bascom and Stose conceived the structure of the Newark basin. As diagrammed in the cross sections of Fig. 9.4 the throw *would* be of the great magnitude mentioned, but if diagrammed as it is in Fig. 9.2, the displacement would be much less.

According to Krynine (1941a) the wedge of sediments is built of coa-

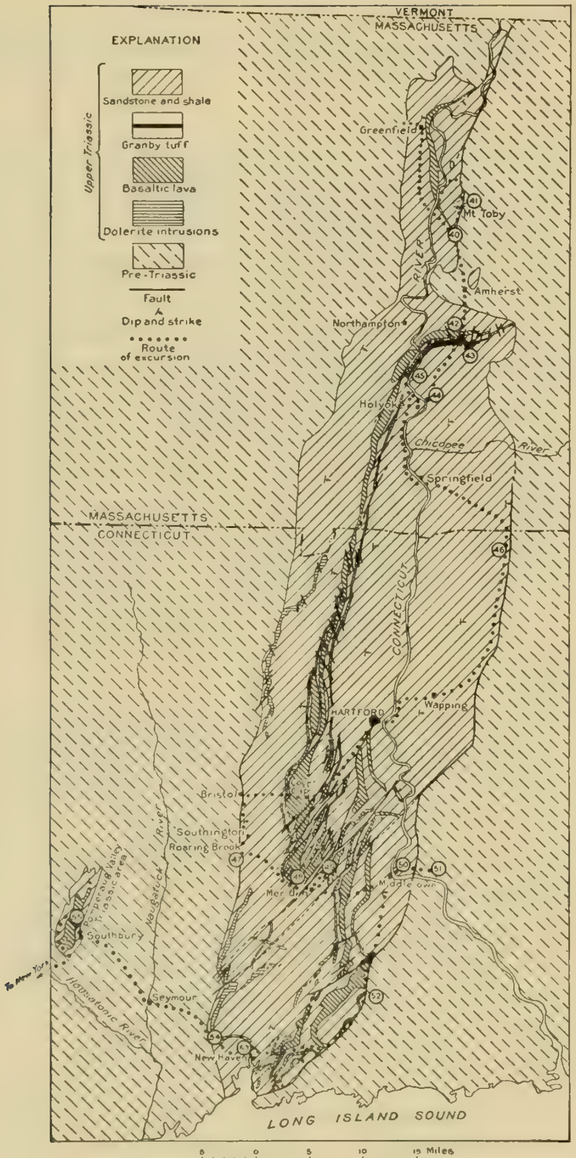


Fig. 9.3. Triassic basin of Connecticut and Massachusetts. Reproduced from Longwell, 1933.

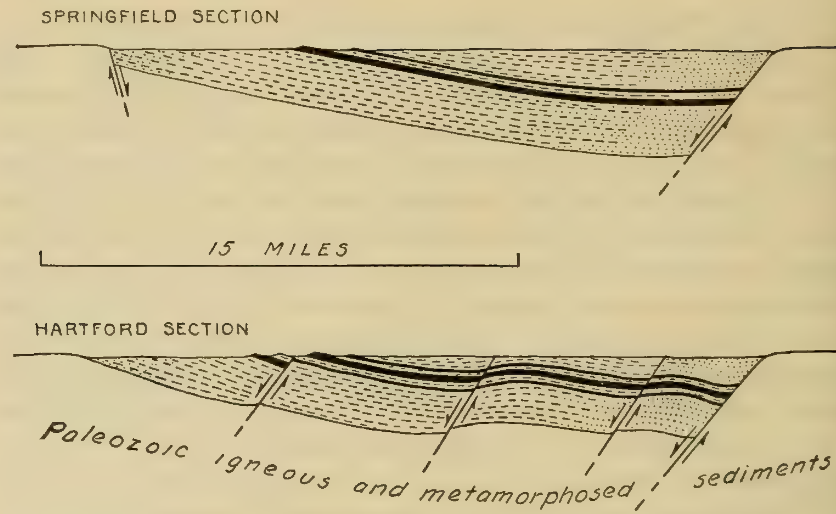


Fig. 9.4. Generalized east-west sections across the Triassic basin in Connecticut and Massachusetts. Somewhat modified after Longwell, 1933.

lescing alluvial fans that radiate westward from the Great Fault and thin from 16,000 to 1500 feet in some 32 miles.

The stratigraphic units are (1) Lower: New Haven arkose, up to 8550 feet, relatively coarse fluvial gray and pink arkoses, conglomerates, red feldspathic sandstones, and subordinate red siltstones and shales; (2) Middle: Meriden formation, up to 2800 feet, fine-grained lacustrine and paludial variegated and dark siltstones, shales, limestones, light feldspathic sandstones, subordinate coarse clastics, and three basaltic lava flows; (3) Upper: Portland formation, up to 4000 feet, fluvial deposit similar to New Haven.

Conglomerates form 10%, sandstones 64%, siltstones and shales 25%, red color is present in 52%. Near the Great Fault sediments pass into fanglomerates.

Two main groups of alluvial fans are present: Central Connecticut (indicolite and little epidote) and Southern Connecticut (no indicolite, much epidote). Almost all the sedimentary detritus is derived from a source area only 5 to 10 miles wide east of the steep, moderately high Great Fault, whose recurrent rejuvenation controlled sedimentation.

Four formations have been mapped on the state geologic map of Massachusetts (1916), but their distribution as continuous-layered units could hardly be shown on cross sections. The central part of the basin at the

surface is marked by the Chicopee shale; this is bordered on both sides and the north by the Longmeadow sandstone, and this in turn by the Sugarloaf Arkose. Along the east side is a coarse border aggregate called the Mount Toby conglomerate. These formations are clearly facies and grade into each other or are interdigitated. The Mount Toby conglomerate is a fanglomerate in large part and an actual talus in others. There can be little doubt about its relation to a great border fault; but in places bedrock crops out surounded by conglomerate, and the position of the fault is obscure.

Intercalated in the clastics and grouped close together in their central part are three lava sheets of diabase. The middle one, the Holyoke diabase, is the thickest and in places reaches 400 feet. Between it and the upper are sandstones that contain large and small reptile tracks which are very well known. Shortly after the third lava outpouring, an explosive eruption took place; and fragments and dust of diabase were spread over a large area to form the Granby tuff. Over the tuff was spread rusty sand in which most of the tracks have been preserved. In the southern part of the basin "dolerite" sheets have been intruded. Dikes are few.

Here as in the other Triassic basins, normal faults cut and displace the beds and volcanic sheets. See sections, Fig. 9.4.

The red color and salt crystal impressions have led a number of writers to envision a semiarid climate; but Krynine, on the other hand, contends that the flora and swamps suggests a precipitation of about 50 inches a year and a temperature of 70° to 80° F. Fresh arkoses and fanglomerates can easily form under tropical humid climate in regions of steep topography. Desiccation marks indicate alternating dry and wet seasons.

STRUCTURE OF BASINS

All the Triassic basins in the eastern United States are bordered on one side or the other by major normal faults. A great fault, although irregular and with branches and perhaps steps borders the Newark basin on the west. The Deep River basin has a major fault on each side. The Connecticut Valley Triassic is bordered on the east by a major fault, also of a complex nature. The long and very narrow basin that stretches from North Carolina into Virginia is bordered on the west by a fault. The several other small and detached basins are shown with faults on either the east or west sides on the *Geologic Map of the United States.*

Associated with all the great border faults and perhaps due to them is a general dip of the beds and sills toward them. See cross sections of Figs. 9.2 and 9.3. The dips range from 5 to 50 degrees and are more generally 10 to 20 degrees. The Triassic beds are not folded as the underlying Paleozoics and metamorphics, upon whose beveled edges they rest unconformably.

Strike faults within the sediments are known, somewhat parallel with the border faults, and many transverse faults cut and offset the beds and sills. In places the transverse faults terminate against normal strike faults and produce a rhombic pattern. Some of the transverse normal faults have been traced out into the folded and thrust-faulted Paleozoic rocks which they also offset.

The normal faults within the basin cut the Triassic sediments and sills, yet some of the dikes associated with the sills follow cross faults. It is generally concluded that the faulting is later than most of the beds, but before the end of the period of volcanic activity, so that most of the sills are cut by the faults, yet some dikes were injected immediately into the fractures when they formed.

ORIGIN OF BASINS

The Triassic basins of the Piedmont province and of the Connecticut Valley have a similar history. The troughs in which the sediments were deposited are due mainly to downfaulting with a major fault or chain of faults on either the outer or inner side. The trough block rotated by settling most adjacent to the border fault. The border faulting is conceived as a fairly continuous process during which the sediments accumulated in the basins as they were progressively deepened. Stose and Bascom (1929) represent sedimentation in the Newark basin to have started considerably before the border faulting began (see Fig. 9.2); then with the onset of faulting the previously deposited beds which came from the southeast were tilted, and the site of later sedimentation, with continued faulting,

shifted more toward the northwest. Also, with the onset of faulting, fanglomerates were washed in by torrential streams from the uplifted block. In the Deep River basin, with faulting on both sides, the fanglomerates came from both directions. If Triassic sedimentation started before faulting, it may have been due to one of two causes: (1) a broad syncline may have developed which later broke into faults on one or both sides, or (2) a change may have occurred from a warm humid climate in whch red soils were developed on the surrounding lands to an arid or semiarid one in which salt crystals developed in the sediments from time to time, and in which torrential floods were common.

The throw of the border faults according to the cross sections of Fig. 9.4 would equal the total thickness of the basin sediments, and therefore, would be of the magnitude of 20,000 feet. This is twice as much as postulated or computed for any other post-Proterozoic normal fault in North America, and leads one to regard the large figure critically. Stose and Bascom (1929) compute the throw at 6000 feet in the southeastern Pennsylvania area by means of their postulated origin of the Newark basin.

The nature of the faults of the Triassic basins, both in vertical and horizontal position and movement, and the general plan of the entire zone of faults from the Carolinas to Nova Scotia reminded Bain (1941) of the Rift Valleys of Africa, and he considers them a rift zone.

LATE TRIASSIC PHASE (PALISADES OROGENY)

The onset of faulting that formed the troughs in which the Triassic sediments accumulated marked the beginning of the Palisades orogeny. It started in late Triassic time and probably ran its course before the end of the period. After the border faults had become major faults and great thicknesses of sediments had accumulated, vast amounts of basic magma entered the basins, chiefly along the border faults, and spread into the sediments as numerous sills, some exceedingly thick, and as great dikes. In places the dikes cut long distances into the country rock. Great amounts of magma reached the surface as basalt flows, which were immediately buried by the accumulating sediments. Accompanying the igneous activity was an additional episode of faulting. Both strike and transverse parallel faults provided avenues of ingress of the magma, and continued faulting broke and offset some of the sills as well as the sediments. The great border faults undoubtedly also continued active in places, dropping the basins farther and inviting new floods of fanglomerate.

The entire activity from the inception of the border faulting through the intrusive and extrusive activity and additional faulting seems to have been fairly continuous and hence not separable into early and late phases. It will all be recognized here as the late Triassic phase, or the Palisades orogeny.

The faulting and dike intrusions spread into rocks adjacent to the Triassic basins, and it is clear that at the time of maximum accumulation the sediments were much more extensive than now. Their beds are beveled on the sides opposite the border faults, and the fanglomerates still bury in places the fault scarps and spread considerable distances over the upthrown blocks. Whitcomb (1942) considers the Spitzenberg conglomerate as a Triassic outlier 20 miles north of the present margin of the Newark basin. The now separate basins may easily have been confluent in places, but such cannot be proved, it seems. It is also possible that, while the Palisades orogeny was taking place in the Piedmont and folded Appalachians, the continental margin lay 100 to 200 miles eastward and Triassic sediments were accumulating there.

10.

ATLANTIC COASTAL PLAIN AND ADJACENT OCEAN BASIN

EXTENT AND CHARACTER OF SEDIMENTS

The Atlantic Coastal Plain is underlain by poorly consolidated Quaternary, Tertiary, and Cretaceous sediments that dip gently seaward. The Cretaceous sediments form a narrow inland belt of outcrop, and the Cenozoic sediments a broad outer belt. In places, the Cenozoic sediments overlap the Cretaceous entirely and rest on the crystalline rocks of the Piedmont. See the *Geologic* and *Tectonic* maps of the United States. The surface is nearly a plain, as the term coastal plain implies. The interruptions to the plain are low, inland-facing questas and, in places, slightly intrenched streams that cross the Cretaceous and Tertiary rocks as they flow toward the Atlantic. The Virginia, Delaware, Maryland, and New Jersey section of the Coastal Plain is one of great estuaries in which tide waters reach across the plain to the Piedmont. These are regarded as drowned river valleys.

The Coastal Plain as a geologic unit extends out into the Atlantic Ocean and forms the broad and well-known continental shelf there. Off Cape Hatteras, the shelf is only 30 miles wide, but both northeastward and southwestward from the cape it broadens. Off New England, it is over 250 miles wide. See the *Tectonic Map of the United States* and Fig. 7.1 of this book.

The Atlantic Coastal Plain is continuous with the Gulf Coastal Plain, which is described in Chapter 41. Florida has been included in the Gulf Coastal Plain, so will not be treated here.

STRATIGRAPHY

The stratigraphy of the Atlantic Coastal Plain is illustrated by a chart, Fig. 10.1 and five cross sections, viz., numbers 32, 33, 34, 35, and 36 of Figs. 10.2, 10.3, and 10.4. Refer to the index map, Fig. 7.1, for the position of the sections. Three of the sections across the Coastal Plain and two of them run lengthwise of it.

The chief elements of the stratigraphy are the Upper Cretaceous, Eocene, and Miocene. Lower Cretaceous beds have been noted in the northern half of the Coastal Plain, and Oligocene beds in the southern part (South Carolina and Georgia). A thin Quaternary cover is fairly extensive in the area between Chesapeake and Delaware bays and in North Carolina. For details of the stratigraphy, see Richards (1945, 1947).

A well in Maryland penetrated 169 feet of dark red, argillaceous sandstone, apparently of Triassic age. See section 36, Fig. 10.4.

STRUCTURE

Coastal Plain

Regional Dip. With few exceptions, the beds dip gently toward the Atlantic. The crystalline floor upon which the sediments rest dips the

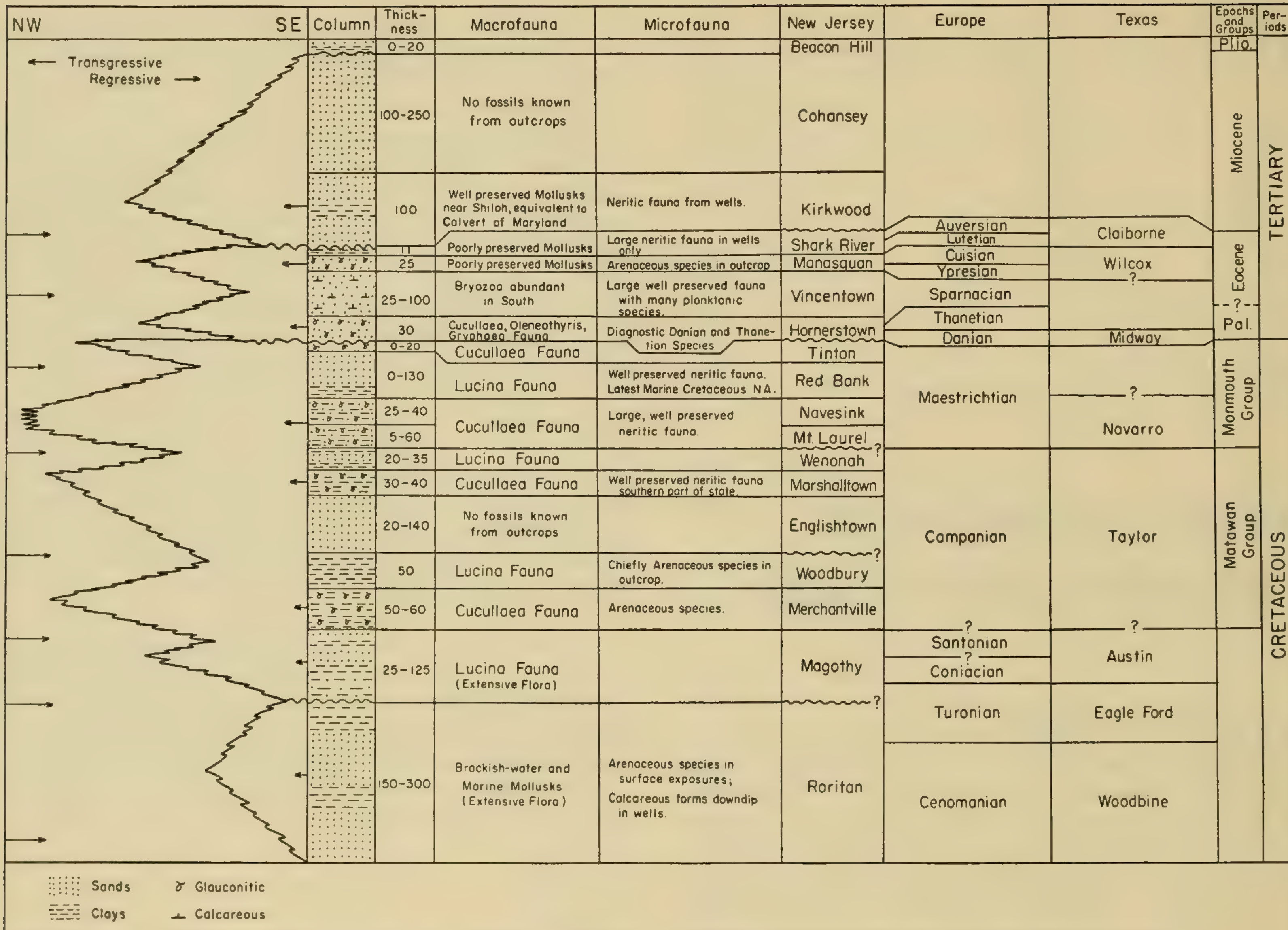


Fig. 10.1. Cretaceous and Tertiary formations in the Coastal Plain of New Jersey. Reproduced from Dorf and Fox, 1957.

greatest amount, because most all the formations thicken seaward, and each successively higher sedimentary surface dips somewhat less than the "basement" floor. From a number of deep wells that have penetrated the crystallines, the ancient surface can be contoured as shown in Fig. 10.6. Its gentlest slope is in North Carolina, where a dip of 10 to 14 feet per mile exists from the inner margin (fall line) to the coast in the southeastern part of the state.

It then breaks seaward into a steeper slope of 122 to 124 feet per mile (Berry, 1948). Two deep wells in northern Maryland demonstrate an offshore dip there of about 100 feet per mile (Balsley *et al.*, 1946), and a uniform slope is indicated. The two slopes in North Carolina are taken to mean two peneplains by Berry (1948), but their local development is puzzling if this theory is true.

Unconformities. The great unconformity at the base of the Cretaceous has already been implied in the discussion of the slope of the surface of the crystallines. This ancient erosion surface, buried by the Cretaceous sediments, has been called the fall zone peneplain. See block diagrams 2 and 3 of Fig. 10.8. Since an outer and sharper slope has recently been defined, the ancient surfaces appear more complicated. It will be discussed further when the continental shelf is considered.

The Lower Cretaceous beds do not crop out anywhere along the Atlantic Coastal Plain; they form a subsurface wedge between the crystalline floor and the Upper Cretaceous. The dashed lines of Fig. 10.6 show the extent and thickness variations of the Lower Cretaceous. It will be seen that the wedge corresponds in position approximately to the outer steeper slope of the crystalline floor. The isopachs should be related to those of Plate 11 which depicts the distribution of Lower Cretaceous strata in the Gulf Coastal Plain and the Caribbean regions. Not enough is known of the Lower Cretaceous and Upper Cretaceous contact to decipher the relations. The Lower Cretaceous Potomac formation is regarded as nonmarine, and the overlying Tuscaloosa as marine (Richards, 1945).

According to Richards' (1945) correlations the Eocene bevels the Upper Cretaceous beds near Asbury, New Jersey (section 32, Fig. 10.2) and rests on the Lower Cretaceous in parts of Virginia (section 33, Fig. 10.3).

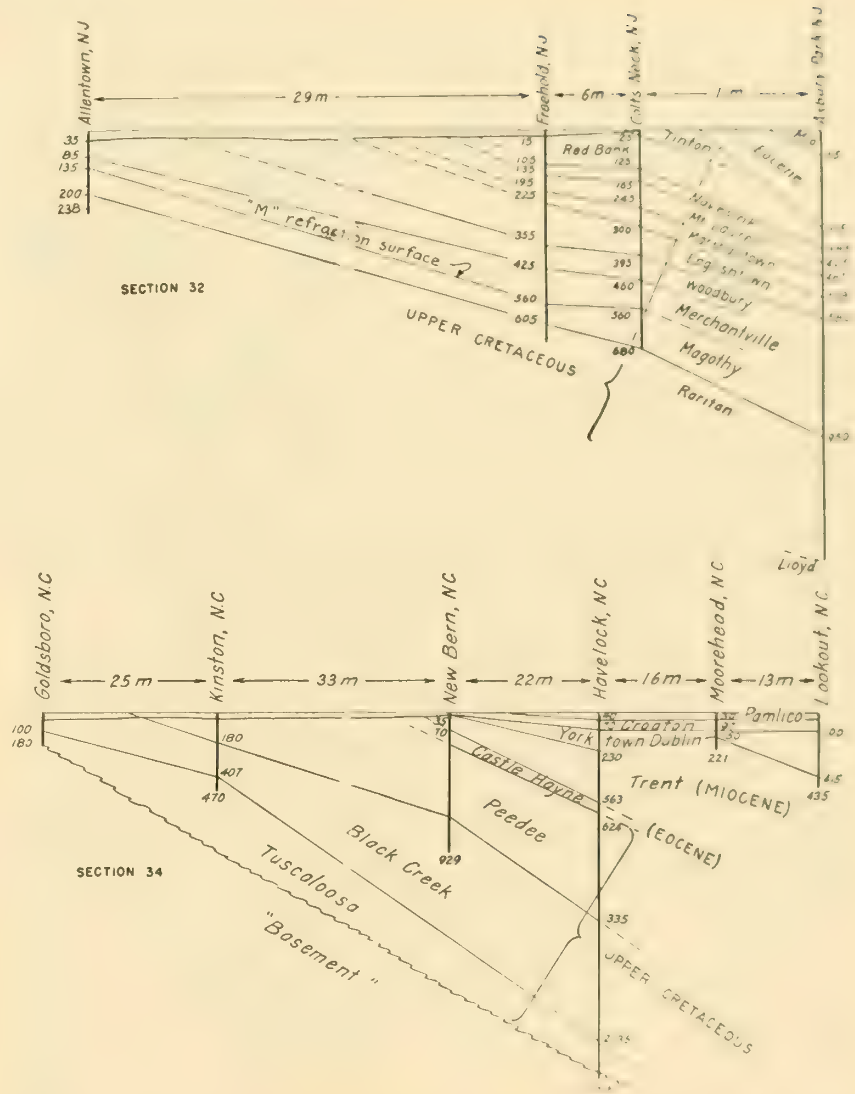


Fig. 10.2. Cross sections of the Atlantic Coastal Plain, after Richards, 1945. Section 32, from Allentown, N. J., to Asbury Park, N. J. Section 34, from Goldsboro, N. C., to Cape Lookout, N. C. See index map, Fig. 7.1, for location of sections.

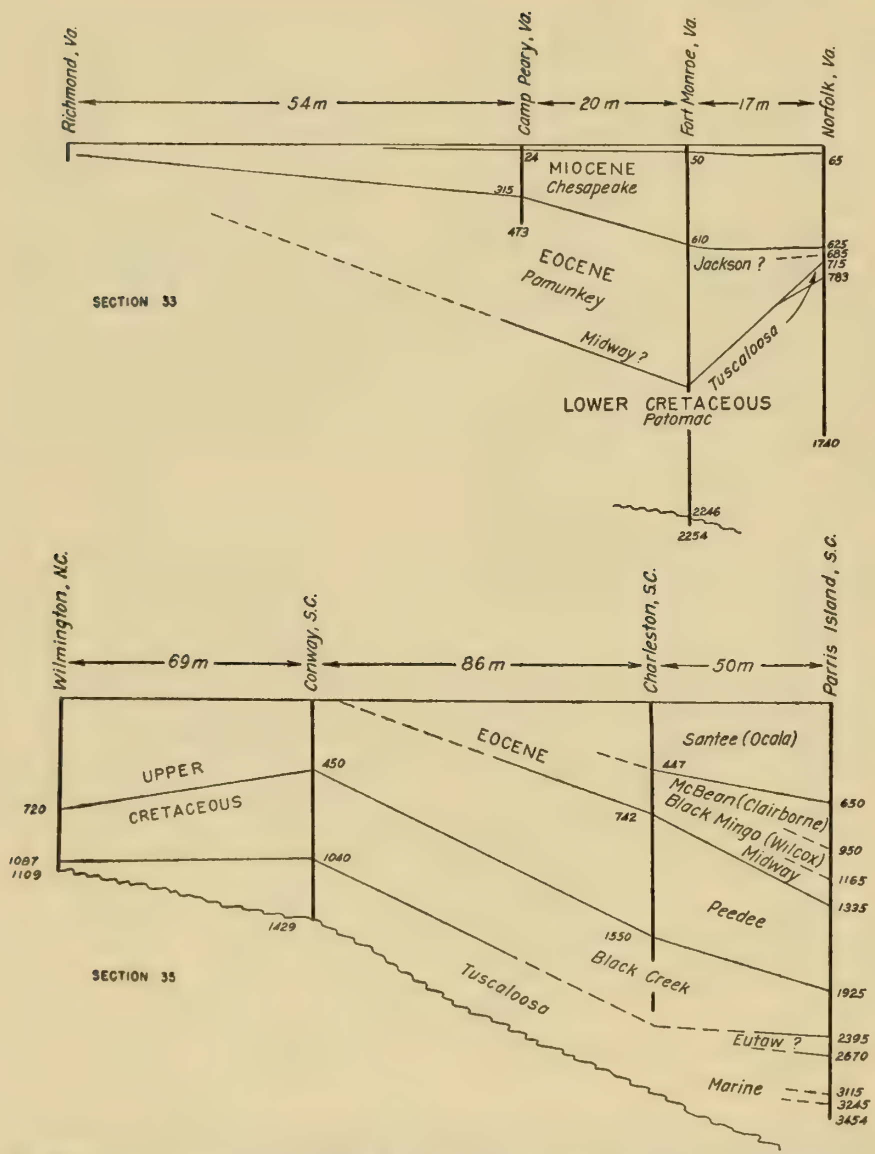


Fig. 10.3. Cross sections of the Atlantic Coastal Plain, after Richards, 1945. Section 33, Richmond, Va., to Norfolk, Va. Section 35, Wilmington, N. C., to Parris Island, S. C. See index map, Fig. 7.1.

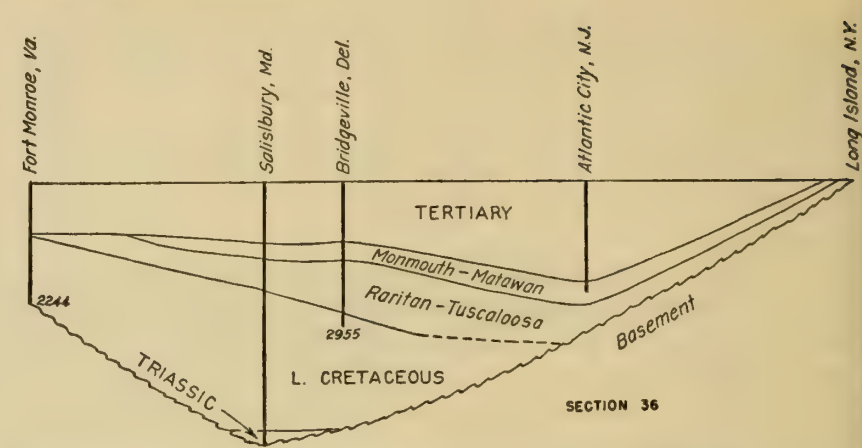


Fig. 10.4. Section of the Atlantic Coastal Plain from Virginia to Long Island, N. Y., after Richards, 1947.

Evidently, therefore, an unconformity of considerable magnitude exists between the Tertiary and Cretaceous systems.

The absence of Oligocene beds, except in the south, suggests an unconformity between the Miocene and Eocene. In most of Richards' sections, however, the Miocene seems conformable on the Eocene. One exception is noted near Summerville, South Carolina. A break, however, occurs between Lower and Upper Miocene in the area between Norfolk, Virginia and Wilmington, North Carolina, where the Yorktown–Duplin formation rests across the entire Lower Miocene, Eocene, and most of the Upper Cretaceous succession. The *Geologic Map of the United States* shows very clearly the unconformity between the Yorktown beds and the entire Upper Cretaceous, Eocene, and Lower Miocene succession in the region adjoining the states of North and South Carolina. Inspection of the map also reveals an unconformity between the Pliocene beds and older ones in this region.

Cape Fear Arch. The most conspicuous feature of the Coastal Plain is the Cape Fear arch of North and South Carolina. See index map, Fig. 7.1, and the *Geologic* and *Tectonic Map of the United States.* Structure contours on the top of the Cretaceous bulge outward at this place and reveal a very broad nose on the regional seaward dip, so the structure is

not truly an arch as defined in Chapter 2. The Eocene and Miocene contacts with the Cretaceous also reflect the broad nose. The unconformities around the Cape Fear arch indicate the principal times of uplift and erosion to have been at the close of the Cretaceous and again at the close of the early Miocene.

In the New Jersey region Dorf and Fox (1957) recognize eight transgressive-regressive cycles of sedimentation in the history of the Coastal Plain from Raritan (Upper Cretaceous) to Cohansey (close of Miocene) time (Fig. 10.1). If these prove to be of local extent, then it would be concluded that the continental margin pulsed up and down locally this many times, but if the cycles are found to be widespread and recorded in the Gulf Coastal Plain sediments, then eustatic changes in sea level would be the more probable cause. The subject will be considered in Chapter 41 on the Gulf Coastal Plain.

CONSTITUTION OF CONTINENTAL SHELF AND ADJACENT ATLANTIC OCEAN CRUST

Composition of Basement

As a result of seismic refraction studies in the Atlantic Coastal Plain between Virginia and New Jersey, Ewing *et al.* (1939) believe that the rocks of the crystalline Piedmont, as known in the exposed belt, are also present in the basement complex below the Cretaceous. They recognize the Petersburg granite and the Wissahickon schist into which the granite is intrusive, under the unconsolidated sediments east of Petersburg, Virginia, and think they can trace the belts northward through Maryland, Delaware, and New Jersey. It would appear, they say, that the Petersburg granite is a correlative of, or is continuous with, the late Devonian granites of Connecticut and Rhode Island.

Deposits of Continental Shelf

The continental shelf off the Atlantic Coastal Plain has been investigated geophysically in the past 12 years, and some interesting results have been obtained (Ewing *et al.*, 1937, 1940). Several seismic traverses were run across the Coastal Plain in order to check the seismic data with known outcrops and well records; and two submarine traverses were run across the continental shelf, one from Woods Hole southward, and one from Cape Henry, Virginia, eastward (section 37 of index map, Fig. 7.1. The Cape Henry section is the most significant. Many reflection surfaces were recorded in the sediments above the crystalline floor, and two particularly strong ones were measured by refraction beyond the present shore line. See Fig. 10.5. The seismic data on the crystalline floor are in fair agreement with the deep-well records and indicate that at a point 60 miles at sea off Cape Henry the basement would be 12,000 feet deep. The significance of the other two surfaces is not altogether clear. Miller (1937) suggests that the "unconsolidated" zone consists of Cenozoic and Cretaceous, and the "semiconsolidated" zone consists of Jurassic and Triassic. The surface separating the two is known as the M zone to the geophysicists, and this has later been considered as a reflection horizon

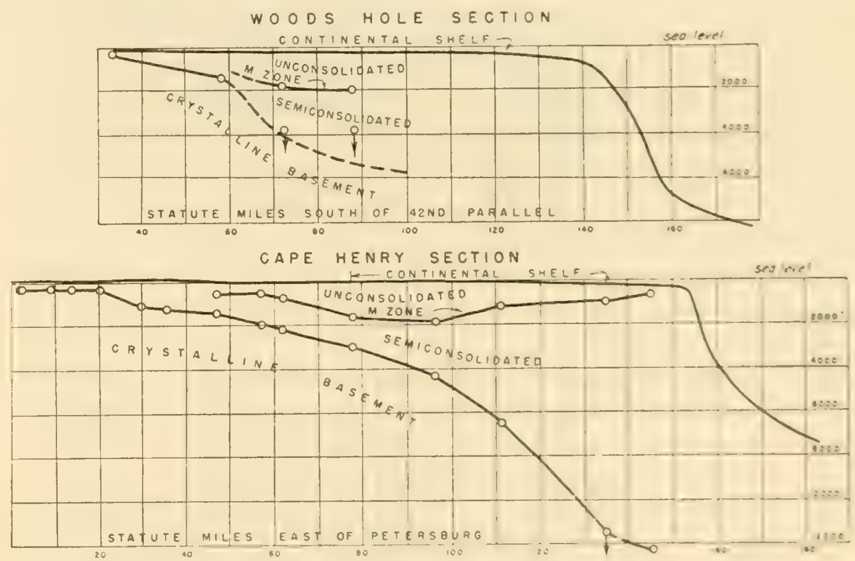


Fig. 10.5. Seismic traverses on the Atlantic Coastal Plain and continental shelf, after Ewing, Crary, and Rutherford, 1937. Small circles represent elevations determined by the refraction seismograph. The Cape Henry section is section 37 of the index map of Fig. 7.1. The Woods Hole section runs southward from Woods Hole, Mass. The M Zone is probably a horizon within the Upper Cretaceous.

in the Upper Cretaceous (Ewing *et al.*, 1939). Richards thinks the M zone in the Upper Cretaceous is the contact between the Magothy and Matawan or Magothy and Merchantville formations. See section 32, Fig. 10.2. If so, about 700 feet of Upper Cretaceous strata, which generally underlie the Magothy, and 1000 feet or more of Lower Cretaceous would be in the semiconsolidated layer. In southeastern Virginia the Eocene rests on the Lower Cretaceous, and the M zone is probably absent; but perhaps seaward the Upper Cretaceous comes in again, and the zone is present.

Contour of Crystalline Basement Surface

In a paper of 1950, Ewing *et al.*, report on profiles off Cape May, New York, and Woods Hole, and concluded that the Precambrian surface does not slope constantly toward the Atlantic Ocean basin floor but has a pronounced reversal of dip at a depth of 16,000 feet before the margin of the shelf is reached off Long Island and Delaware Bay. Structure contours on the surface are shown on Fig. 10.6. Farther south off the Cape Fear arch the slope of the crystalline floor reflects the arch nearly to the margin of the shelf (Richards, 1945, 1947; Berry, 1948; Hersey *et al.*, 1959). The surface is lost seaward over the Blake Plateau, where no seismic record of it or deeper boundaries of velocity layers were obtained. See Fig. 10.7. The strike of the surface veers westward in South Carolina and northern Florida. Near Jackson, Florida, the surface dips steeply southward and is lost at a depth of 19,000 feet. The basement contours here are distinctly discordant to contours drawn on the top of the Cretaceous (Fig. 10.6).

In the shelf profiles off Long Island and Delaware Bay the unconsolidated sediments thicken gradually outward under the shelf. In the upper section of Fig. 10.7 Heezen *et al.* (1959) show a ridge of basement rock at the shelf margin and then an abrupt fall-off apparently of fault nature. Oceanward is a second basin in which the unconsolidated and consolidated sediments attain a maximum thickness of 33,000 feet (10.3 kilometers). The unconsolidated layer thins over the deep Atlantic floor to about 2 kilometers, but becomes much thicker again on the approaches to the Bermuda Rise and Mid-Oceanic Ridge.

The seismic profiles across the Atlantic Coastal Plain and continental shelf to date have been summarized by Drake *et al.* (1960), and these writers point out that a ridge of basement rock near the edge of the shelf is a common feature. It separates two sedimentary troughs, one under the shelf, and another in deeper water under the shelf slope and rise. The ridge and basins can be seen in the upper section of Fig. 10.7 and section A–A′ of Fig. 11.34. The sediments in the inner or shelf trough have been drilled in several places along the Atlantic Coastal Plain and are mostly shallow water sands, silts, and clays. Cores of the upper part of the sediments of the outer trough have revealed features attributed to slumping, sliding, and turbidity currents, and are in part similar to graywackes. Drake *et al.* point out that the size of the troughs and the thickness and character of sediments in them are similar to the early Paleozoic troughs of the Appalachians as restored by Kay (1951) and that here is a good representation of the miogeosyncline (inner trough) and eugeosyncline (outer trough). Compare with Figs. 11.17, 6.6, and 6.15. Evidence of past volcanism in the outer trough is present in the form of partially buried seamounts with large magnetic anomalies.

The eugeosyncline, according to the above view, develops largely on the oceanic crust, and represents, when uplifted, an accretion to the continent. The theory appears very attractive when related to the Paleozoic Cordilleran geosyncline.

Submarine Canyons. Comprehensive submarine surveys of the whole of the continental shelf and slope of the northeastern United States have been made since 1930, using the most advanced methods, and the results were published in 1939. Chart 1 of the publication, "Atlantic submarine valleys of the United States and the Congo submarine valley" (Veatch and Smith, 1939) is a composite of all the modern work from Cape Hatteras to Georges Bank. The same results are presented in reduced scale and somewhat simplified on the *Tectonic Map of the United States.*

The shelf is a fairly smooth plain and a continuation of the emerged Coastal Plain. The most prominent feature is the Hudson channel, which is entrenched 50 to 150 feet in the shelf from the mouth of the Hudson River to near the edge of the shelf. South of the channel are many shallow depressions and low, irregular ridges trending generally parallel with the shore. They have been likened to bars and lagoons. Northeast of the channel, the shelf is a regular oceanward slope, perhaps rilled with many

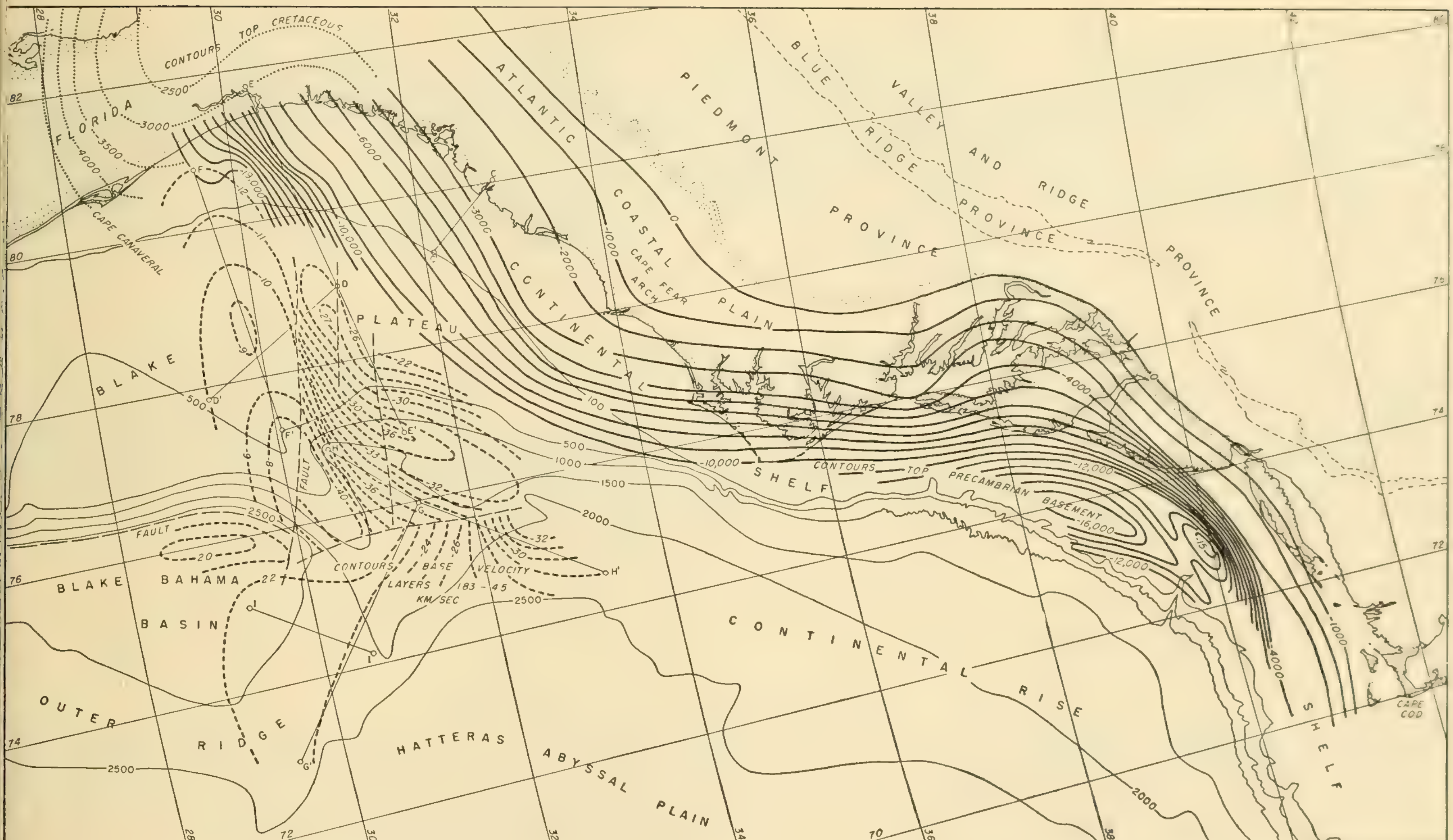


Fig. 10.6. Structure of Atlantic Coastal Plain and adjacent ocean crust from Cape Canaveral to Cape Cod. Thin contour lines are on the ocean floor and are in fathoms. The following contours are in feet. Heavy continuous lines are structure contours on the top of the Precambrian crystalline basement. Heavy dashed lines are on the base of velocity layers interpreted to be sediments. Dotted contours are on top of Cretaceous. Compiled from Ewing *et al.*, 1950 and Hersey *et al.*, 1959.

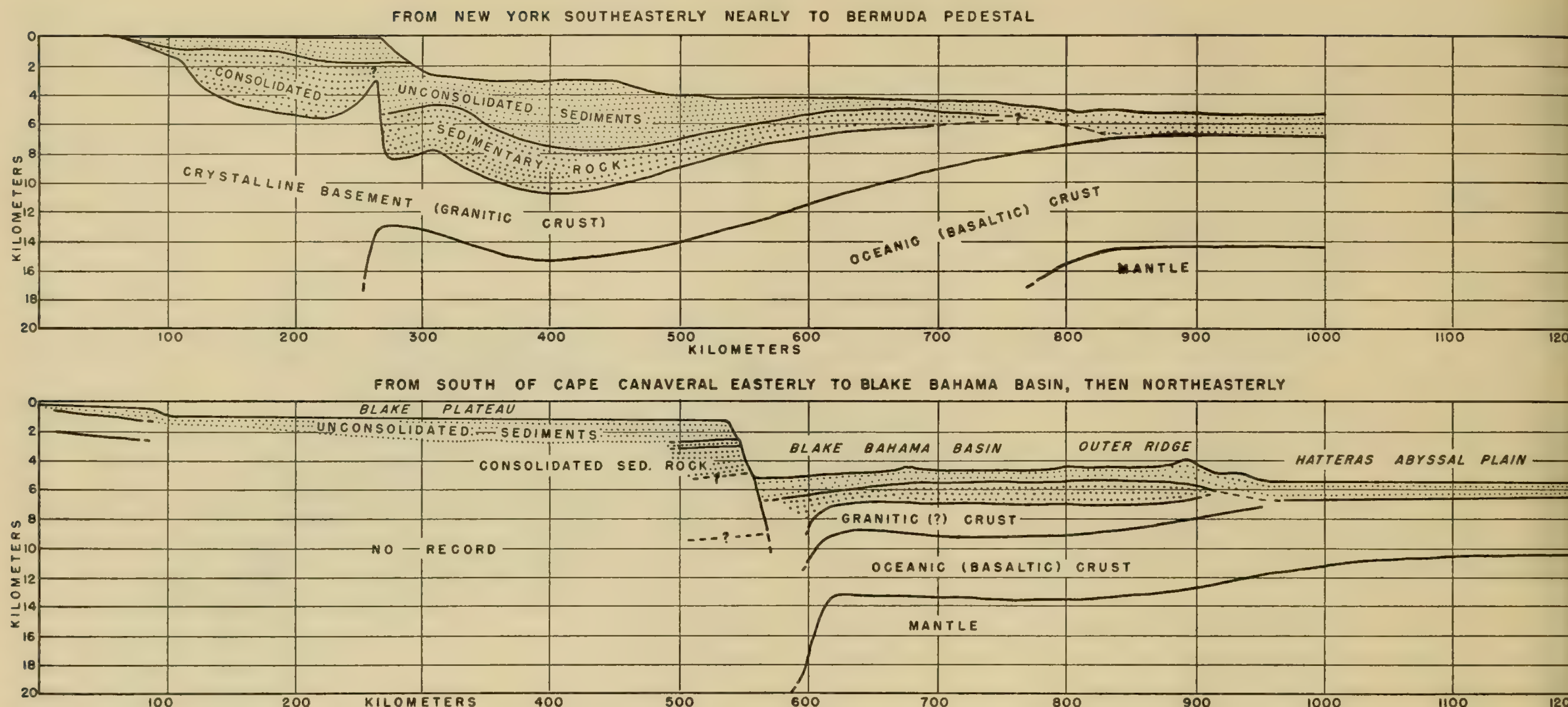
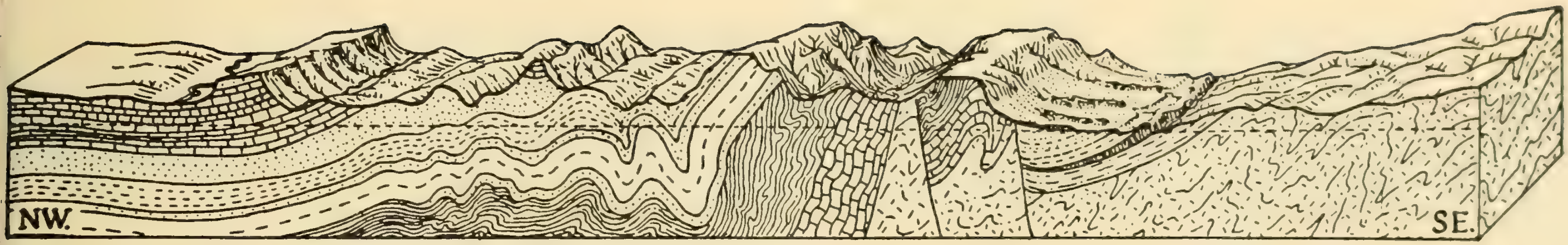


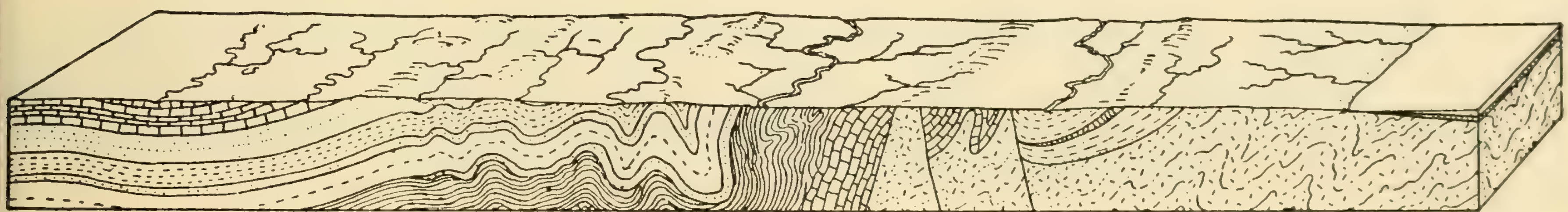
Fig. 10.7. Sections of the crust of the Atlantic Coastal Plain and adjacent ocean.

shallow valleys. The shelf breaks abruptly at about the 600-foot depth to a steeper slope, known as the "slope" which carries down to 8000 feet and more below sea level in approximately 50 miles. In a few sections, the slope is as steep as 700 feet per mile (7½ degrees).

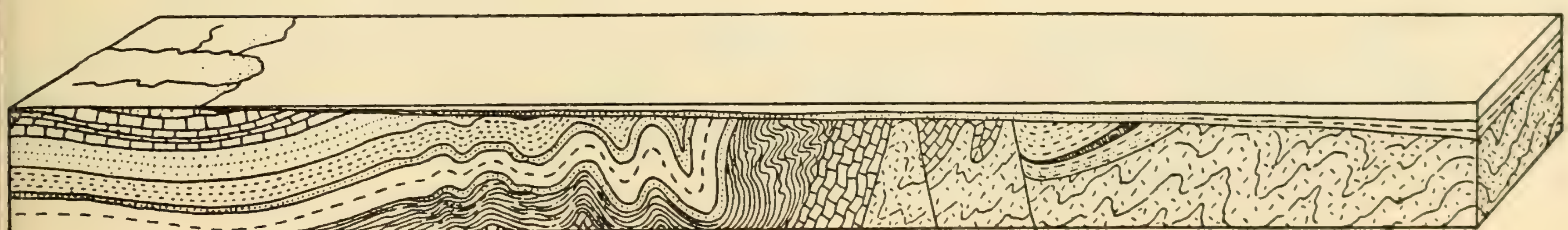
The slope is riven by two kinds of dip-slope features; canyons that extend headward into the shelf 10 to 30 miles from the outer margin, and numerous deep parallel rills that are limited entirely to the slope. The bottoms of the submarine canyons range from 2000 to 3700 feet below the floor of the shelf at the outer margin. Those south of the submarine Hudson channel and canyon generally lose their identity on the slope, merging with the many rill-like canyons or not being larger than the canyons limited to the slope. The Hudson canyon and others that indent the shelf to the eastward along Georges Bank more clearly retain their identity down the slope. Only one submarine canyon in this section of continental shelf can be related with any assurance to a major river on land. This singular relation is the Hudson, whose channel from New York has been mapped about 100 statute miles across the shelf to the head of the deep shelf-indenting and slope canyon.



Rejuvenated Appalachians in post-Newark time

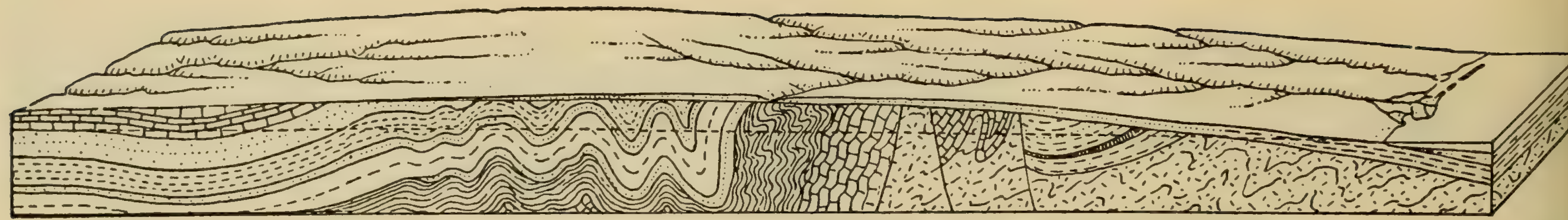


The Fall Zone peneplain



Encroachment of Cretaceous sea and deposition of Coastal Plain beds

Fig. 10.8. Early stages in Appalachian epeirogeny. Reproduced from Johnson, 1931. Diagrams 1, 2, and 3 from top to bottom.

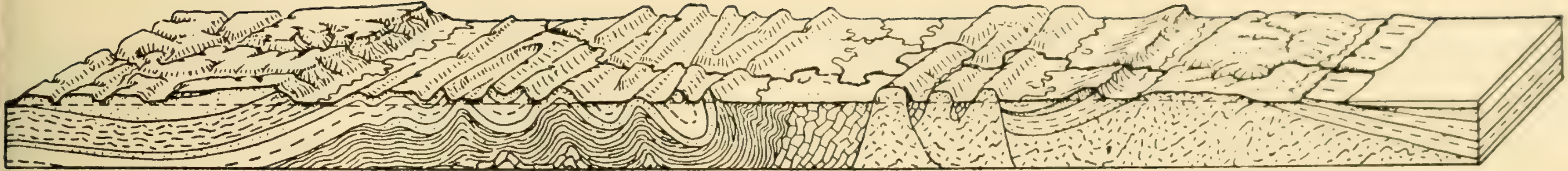


Arching of Fall Zone peneplain and its Coastal Plain cover; regional superposition of southeastward-flowing streams

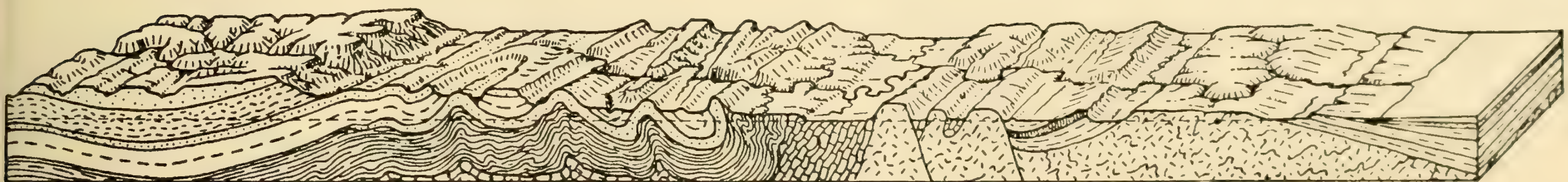
The Schooley peneplain

Arching of Schooley peneplain

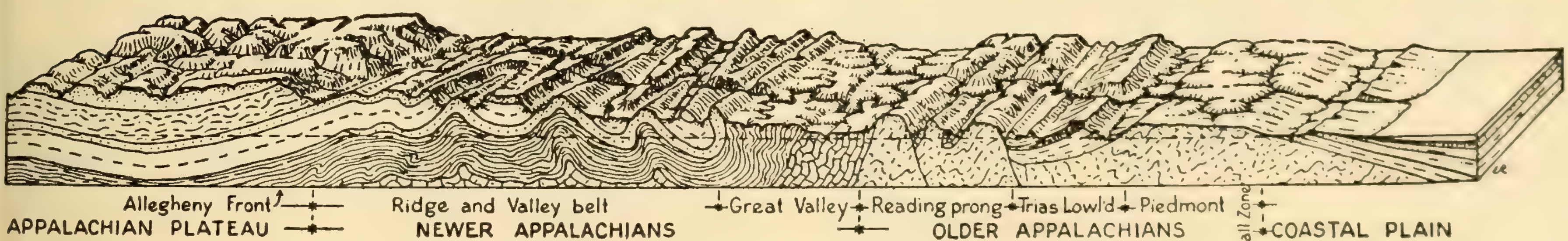
Fig. 10.9. Tertiary stages in Appalachian epeirogeny. Reproduced from Johnson, 1931. Diagrams 4, 5, and 6 from top to bottom.



Dissection of Schooley peneplain and development of Harrisburg peneplain on belts of nonresistant rock



Uplift and dissection of Harrisburg peneplain and development of Somerville peneplain on the weakest rock belts



Uplift and dissection of Somerville peneplain to give present conditions

Fig. 10.10. Late Tertiary and Quaternary stages of epeirogeny and erosion in the Appalachians. Reproduced from Johnson, 1931. Diagrams 7, 8, and 9 from top to bottom.

Fig. 10.11. Physiographic provinces, Atlantic Ocean. Reproduced from Heezen *et al.*, 1959.

The Atlantic continental shelf is most probably constructional and due to sedimentation influenced by fluctuating sea level during the Pleistocene. Although certain early writers during a vigorous controvery (1930–1940) contended that the canyons are due to subaerial erosion, and therefore that the Atlantic coast has subsided 5000–10,000 feet subsequently, the theory is generally held today that the canyons are due to submarine slumping, mud flows, and turbidity currents. See discussion in Chapter 32 of submarine canyons off the California coast.

Appalachian Epeirogeny

Following the Appalachian orogeny in the late Paleozoic and the Palisades orogeny in the late Triassic, a long period of erosion set in and

lasted during all of the Jurassic. By the beginning of Cretaceous time, an extensive and very subdued surface across the folded and thrust-faulted Appalachians, and across the Blue Ridge, the Triassic basins, and the Piedmont had formed. This is known as the fall zone peneplain. Study diagrams 1 and 2 of Fig. 10.8. The entire area as far westward as the plateaus province, according to Johnson (1931), was then invaded by shallow epicontinental seas, and in them Cretaceous sediments were deposited (diagram 3). From subsurface studies of the Coastal Plain sediments, it has been shown that the Lower Cretaceous is entirely buried by the Upper, and it appears that the extensive overlap that Johnson visualizes occurred in Upper Cretaceous time. Others admit that the Cretaceous extended farther inland than the present erosional margin but do not believe that it extended beyond the Blue Ridge. Johnson and later Strahler (1945) believe the overlap necessary to explain the stream pattern of the Ridge and Valley province.

The fall zone peneplain was then arched broadly with the crest in the Ridge and Valley and Blue Ridge provinces and the flanks far westward in the plateaus and far eastward in the site of the present Coastal Plain and continental shelf. Another episode of base-leveling followed, which, like the previous one, established an extensively subdued surface, but lower and younger. This is known as the Schooley peneplain. See diagrams 4 and 5 of Fig. 10.9. The only remnant of the fall zone peneplain is that buried beneath the Cretaceous sediments of the Coastal Plain. The Schooley surface is now generally recognized in remnants as the highest flat tops of ridges in the Appalachian region.

Broad arching again occurred, and the Schooley peneplain was dissected in the manner represented in diagrams 6 of Fig. 10.9 and 7 of Fig. 10.10. A few master streams persisted across the folds and thrusts, while many subsequent streams etched out the resistant formations to produce the first appearance of flat-topped, subparallel, ridges and valleys. The new base level below the flat-topped ridges is known as the Harrisburg peneplain. See diagram 7 of Fig. 10.10. Still third and fourth stages of arching are recognized in the dissection of the Harrisburg peneplain and the establishment of the lower Somerville surface, and the dissection of the Somerville to the present stream bottoms. See diagrams 8 and 9 of Fig. 10.10. An extensive literature may be found on the geomorphology of the Appalachians, and most premises and conclusions of the above summary of Johnson's work have been contested. Most authorities recognize the vertical uplift, but some contend that a symmetrical arching did not occur. It may also be argued that the arching was a slow, continuous process, and not one of four stages with interims of standstill.

Physiographic Provinces of North Atlantic Floor

Echo sound tracts of fifty expeditions in the North Atlantic including over 200,000 miles by vessels of the Lamont Geological Observatory with the Luskin precision depth recorder were compiled by Heezen *et al.* (1959), and a physiographic relief map of the ocean floor was prepared. From it the physiographic provinces are resolved as shown on the map of Fig. 10.11. Profiles to accompany the map are reproduced in Fig. 10.12. There are three major divisions, each with its subdivisions as follows:

Continental Margin
- *Category I*
  - Continental Shelf
  - Epicontinental Seas
  - Marginal Plateaus
- *Category II*
  - Continental Slope
  - Marginal Escarpments
  - Landward Slopes of Trenches
- *Category III*
  - Continental Rise
  - Marginal Trench–Outer Ridge Complex
  - Marginal Basin–Outer Ridge Complex

Ocean Basin Floor
- *Abyssal floor*
  - Abyssal Plains
  - Abyssal Hills
  - Abyssal Gaps and Mid-Ocean Canyons
- *Oceanic Rises*
- *Seamount Groups*

Mid-Oceanic Ridge
- *Crest Provinces*
  - Rift Valley
  - Rift Mountains
  - High Fractured Plateau

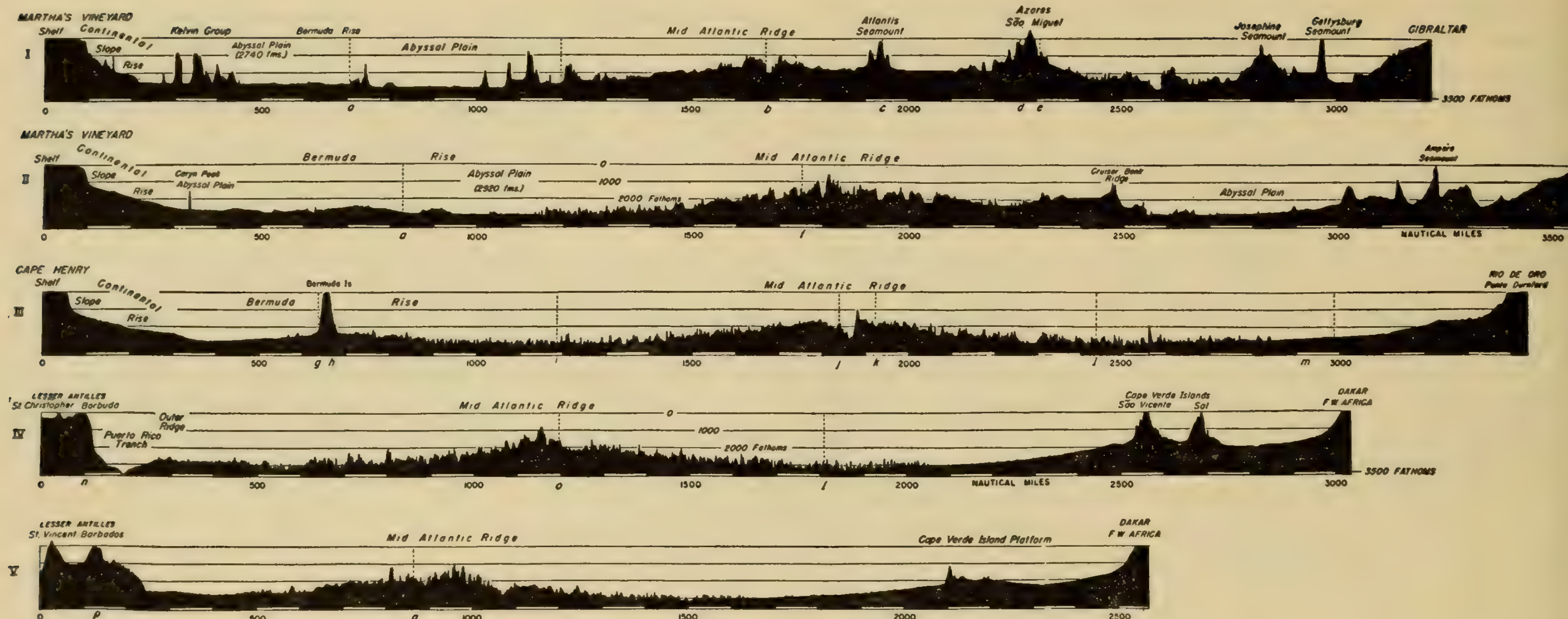


Fig. 10.12. Relief profiles across the Atlantic. Reproduced from Heezen *et al.*, 1959. Letters a to q indicate where sounding profiles of different cruises were joined.

Flank Provinces
Upper Step
Middle Step
Lower Step

For further information the reader is referred to the work of Heezen *et al., Geological Society of America Special Papers 65.* A tectonic map to supplement the publication is yet to appear, but the gross details as now conceived by Heezen and colleagues of Lamont Geological Laboratory are portrayed in the cross section of Fig. 10.13.

Blake Plateau, Blake Bahama Basin, and Outer Ridge

As shown on Fig. 10.11 the continental shelf breaks into two steps south of Cape Hatteras, and the lower step is known as the Blake Plateau. East of the Blake Plateau is the Blake Bahama basin, and east and north of it is the low Outer Ridge. The Outer Ridge swings northwestward at 29° N. Lat., 73° W. Long., and heads toward the Cape Fear arch to merge with the Blake Plateau. Details are given on Fig. 10.6. The outer escarpment of the Blake Plateau is probably a fault scarp, according to Heezen *et al.* (1959). See lower diagram of Fig. 10.7.

A seismic refraction survey of part of the Blake Plateau was made by Hersey *et al.* (1959), and the principal profiles are shown on Fig. 10.6. The same letter designations are retained for the profiles as in the original article. The purpose of the study was to determine the relation of the Plateau crust to the continental crust on one side and to the oceanic crust on the other. Four characteristic profile sections are shown in Figs. 10.14 and 10.15.

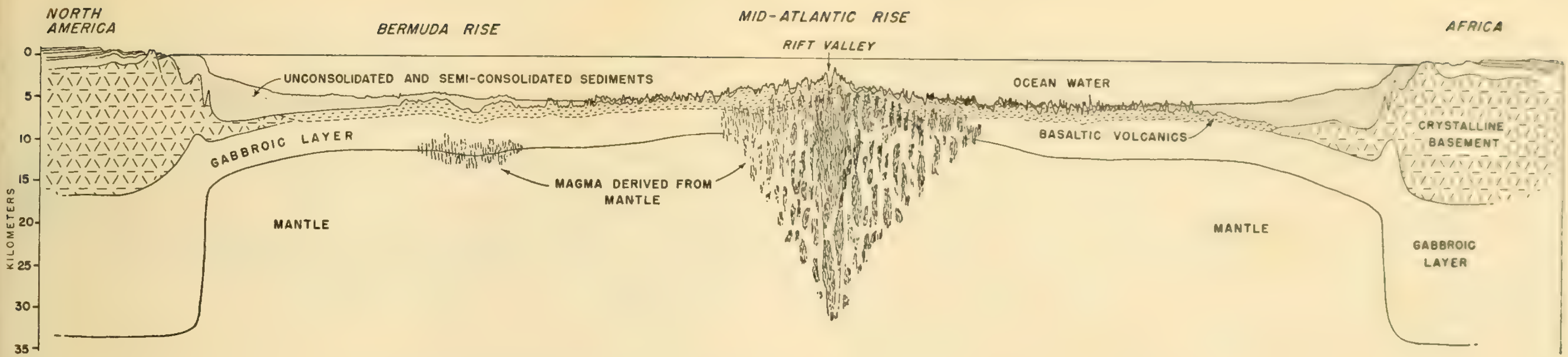


Fig. 10.13. Crustal structure across North Atlantic. After Heezen *et al.*, 1959, with minor changes taken from new section furnished by Tharp and Heezen.

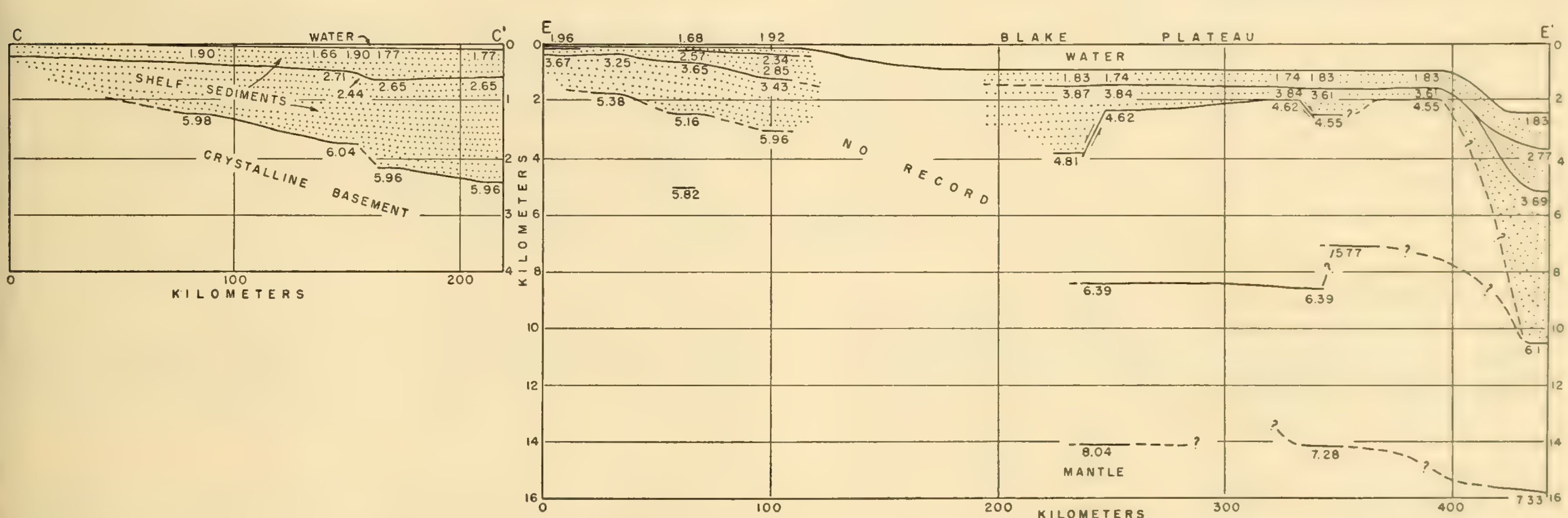


Fig. 10.14. Crustal structure sections C–C′ and E–E′ of Fig. 10.6. After Hersey *et al.*, 1959. Numbers are velocities per second of the various layers. Stippled layers are interpreted as unconsolidated and consolidated sediments.

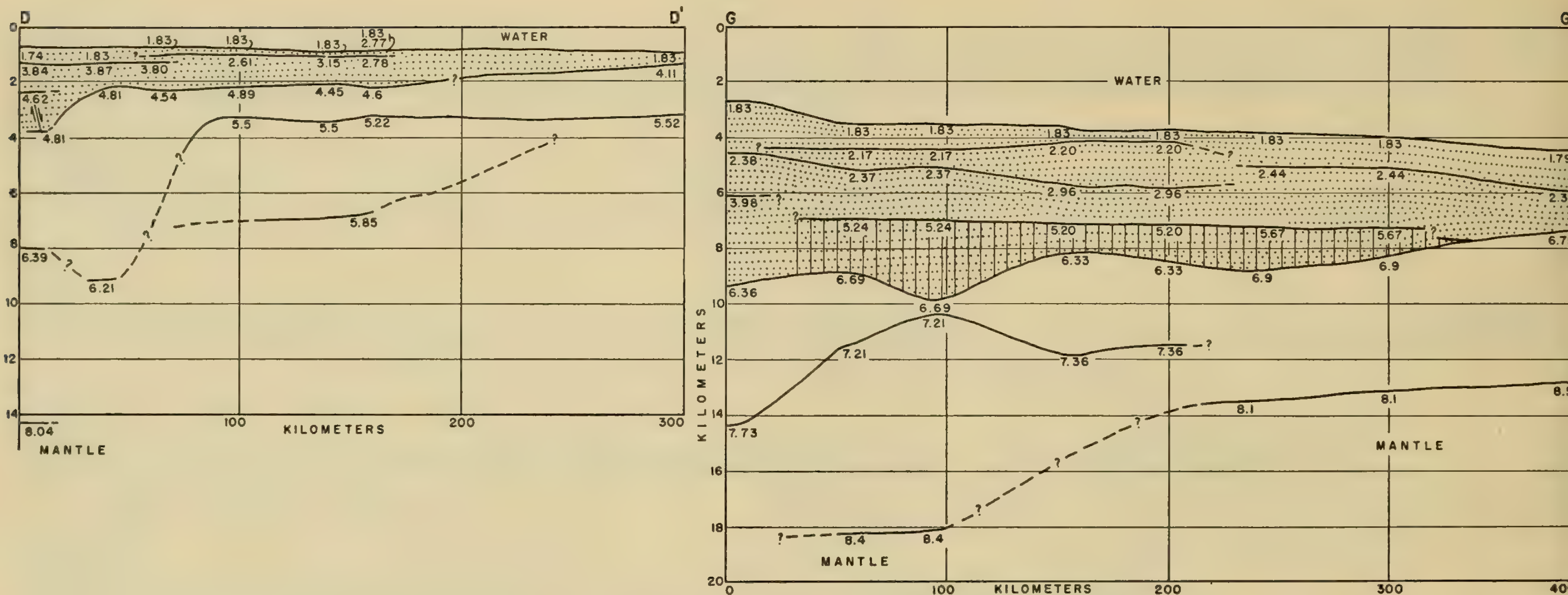


Fig. 10.15. Crustal structure sections D–D′ and G–G′ of Fig. 10.6. After Hersey *et al.*, 1959. Numbers are velocities in kilometers per second. Stippled layers are interpreted as unconsolidated and consolidated sediments.

The results on the continental shelf are correlated with adjacent continental geology. The deepest horizon traced along the shelf is interpreted as granitic basement, which has compressional velocities of 5.82–6.1 km/sec. At the southern extremity it is at a depth of 6 km., shoals to 0.86 km. near Cape Fear, and deepens north of Cape Hatteras to more than 3 km. North of Charleston, South Carolina, there is excellent depth correlation with granitic basement in coastal wells; to the south all deep wells are inland. Age correlations are based on well data near the coast, which indicate to us that most of the observed section is Cretaceous.

On the Blake Plateau, several layers (1.83–4.5 km/sec.) are interpreted as sedimentary. A 5.5–km/sec. layer is found only south of a line from 30°30′ N., 78°W. to Cape Canaveral. Velocities higher than 5.5 km/sec. have been measured on six profiles on the Blake Plateau. The 5.5–km/sec. layer and a 6.2–km/sec. layer appear to form a positive feature to the south of the abovementioned line [indicated as fault on Fig. 10.6]. Higher velocities, 8.0 km/sec., and 7.28 and 7.3 km/sec., which are probably not the same horizon, are found at markedly different depths. Possibly these represent the M layer and ultrabasic material, depending on relations not now known.

[The Outer Ridge along section G–G′] is underlain by thick low-velocity layers (1.83–2.96 km/sec.), interpreted as sediments, and higher-velocity layers which form a distinct linear structure having the same general trend as the ridge. At its northwestern end this trend treminates against a thick lower-velocity section interpreted as a sediment-filled trough (Hersey *et al.*, 1959, p. 1).

An attempt is made on Fig. 10.6 to contour the base of the interpreted sedimentary layers (velocities less than 4.5 km/sec) from the profiles of Hersey *et al.* The results are to be taken simply as pictorial. There seems little doubt, however, that a major fault transects the Blake Plateau, but of a date preceding the deposition of the upper two velocity layers of sediments, because they bury the escarpment. This fault, extended south-

easterly, probably forms the south boundary of the Outer Ridge, described above by Hersey *et al.*, but if so, it does not show in section H–H′. The Blake Plateau south of the fault, at any rate, stands 15,000–30,000 feet above the block on the north, in reference to the base of the interpreted sedimentaries. The deeply filled block extends northward at least to 32° N. Lat.

Regarding the origin of the Outer Ridge, Hersey *et al.* point out that two velocity layers appear there that are unusual, namely the 5.20–5.67-km/sec layer and the 7.21–7.73-km/sec layer. The 5.2-km/sec layer is interposed between the sedimentary layers and the basaltic "oceanic layer" (6.5 ± km/sec), and the 7.5-km/sec layer interposed between the oceanic layer and the mantle. Since profile D–D′ shows a 5.22–5.52-km/sec layer under the Blake Plateau the rock represented by this velocity range is probably not unique to the Outer Ridge. The 7.5-km/sec layer, however, seems more restricted to the Ridge, but it, nevertheless, is known to extend as far north as the northern end of profile H–H′.

The 5.2-km/sec layer is regarded as a mass of extruded volcanic material, lighter and more porous than the basaltic "oceanic crust" layer, and the 7.5-km/sec layer is taken to be a mixture of mantle rock with the oceanic crust, probably by intrusion of peridotitic magma into basalt, in the manner postulated for the Mid-Atlantic Ridge (Fig. 10.13).

Hersey *et al.* (1959) speculate that the ultrabasic intrusions fed the volcanic extrusions, then at the surface, and that the two are complementary. Another theory might be one in which basalt is formed by partial melting of the mantle, with the basalt rising to concentrate in mesh fashion in the upper part of the mantle. This basalt could then rise in fissures and vents through the oceanic crust to eruption at the surface. See Chapter 33 on igneous rock provinces.

Mid-Atlantic Ridge

Topography. The Mid-Atlantic Ridge is a broad arch or swell that occupies approximately the center third of the ocean (Figs. 10.11 and 10.12). The higher and central part is less than 1600 fathoms below sea level, and the flanks fall between 1600 and 2500 fathoms. The Ridge is very rough as the profiles indicate, and the most striking feature is a deep notch or cleft in the crest of the arch, called the Rift Valley. On an average profile the floor of the valley lies at about 2000 fathoms below sea level, whereas the adjacent peaks average about 1000 fathoms. The relief from floor to adjacent peaks ranges from 700 to 2100 fathoms. The width of the valley between crests of the adjacent peaks ranges between 15 and 30 miles; at an elevation of 500 fathoms above its floor the width is from 5 to 22 miles (Heezen *et al.*, 1959).

On either side of the Rift Valley are terranes of sharp and strong relief called the Rift Mountains. Immediately adjacent to the central Rift Valley are the High Fractured Plateaus with local relief of 400 fathoms and ranges 8 to 20 miles apart. Flanking the High Fractured Plateaus is a succession of provinces known as the Upper Step, Middle Step, and the Lower Step. The topography here likewise is rough with local relief of 200 fathoms. Peaks over 200 fathoms high occur at about the frequency of 7 per each 100 miles. The steps appear to be separated from each other by scarps of considerable length.

Seismicity. The High Fractured Plateaus and Rift Valley make up a zone of considerable seismicity. See Fig. 10.16. Another zone extends from the Rift Valley through the Azores eastward to Gibraltar.

Sediments. Photos taken on the sides of seamounts in the Rift Mountains show scour and ripple marks indicating deep-ocean currents. Cores taken in intermontane basins show interlayering as turbidity current deposits.

Rocks. The lithology of the Mid-Atlanic Ridge is known from three sources: (1) rocks dredged from the sea floor, (2) detrital rock fragments found in sediment cores, and (3) rocks exposed on the islands of the Ridge. These all point to olivine gabbro, serpentine, basalt, and diabase as the predominating rock types. One limestone sample probably of Tertiary age was collected from the Rift Valley at about 30° N. Lat. (Heezen *et al.*, 1959).

Crustal Structure. Seismic refraction records have been obtained in about twenty places on the Mid-Atlantic Ridge, and the following layering is reported (Heezen *et al.*, 1959). See Fig. 10.13.

> . . . the average crustal structure of the crest provinces and Upper Step consists of 0.4 km of low-velocity sediment and 2.8 km of rock with a velocity of 5.1 km/sec. overlying a substratum in which the velocity is 7.3 km/sec. The thickness of the layer of low-velocity sediment varies considerably from

Fig. 10.16. Earthquake epicenters, North Atlantic. Reproduced from Heezen *et al.*, 1959.

place to place. In the crest provinces the 5.1 km/sec layer is commonly exposed. In the flank provinces appreciable thicknesses (to 1 km) of sediment have been measured.

Under the abyssal floor of the ocean the low velocity sediment layer is underlain by a 6.7-km/sec layer, and this by a 8.1-km/sec layer. The lower is considered the mantle of peridotite and the overlying layer a gabbroic or firm basalt layer. Under the Ridge neither of these two are present but instead layers of 5.1-km/sec and 7.3-km/sec.

Ewing and Ewing (in press) suggest that this intermediate velocity (7.3 km/sec.) is the result of a physical mixture of oceanic crustal rocks and mantle rocks. To explain such large-scale mixing they propose that extensive vulcanism and intrusion along the Mid-Atlantic Ridge have produced an

intermingling of the crustal and mantle rocks, and that this was associated with convection cells in the deep mantle which supply large quantities of basaltic magma and produce extensional forces on the crust and upper mantle (Heezen *et al.*, 1959).

In a paper (in press) Heezen and Ewing compare in detail the topography and seismicity of the African rift valleys and the Rift Valley of the Mid-Atlantic Ridge. Their conclusion is that the two areas are of basically the same structure, and in fact both form parts of the same continuous structural feature. Since the African rift valleys seem clearly to be the result of normal faulting resulting from extension of the crust, Heezen and Ewing conclude that the topography of the Mid-Atlantic Ridge is largely the result of normal faulting. Whether the forces are the result of horizontal extension or vertical uplift remains the most important unsolved problem in connection with the origin of the continental as well as the sub-oceanic rift-valley systems. Hess (1954) has proposed a mechanism relating suboceanic uplift to expansion due to serpentization of the upper mantle (Heezen *et al.*, 1959).

11.

NEW ENGLAND APPALACHIAN SYSTEMS

DIVISIONS OF NEW ENGLAND APPALACHIANS

The New England Appalachian systems will be divided for purposes of discussion into a western belt and an eastern. The western belt includes those structures in and on either side of the Hudson Valley and Lake Champlain lowlands, and the eastern belt includes a north-south zone through central and eastern Vermont, New Hampshire, and Maine. The western zone is essentially the core of the Late Ordovician Taconic orogeny and the eastern the site of the Late Devonian Acadian orogeny. A third division may be recognized through Rhode Island and Massachusetts on the far east where Carboniferous basins and related igneous activity indicate a still later orogenic belt.

HUDSON VALLEY LAKE CHAMPLAIN REGION

Relief Features

The relief features of the Taconic orogenic system stretch along the general Hudson Valley, Lake Champlain lowlands, and St. Lawrence Valley. In addition to hills and ridges within the lowland, it is convenient under this heading to discuss the Hudson highland and Catskill and Adirondack Mountains on the west, the Laurentian highlands on the northwest, and the Taconic and Green Mountains on the east. The Taconic orogeny culminated in late Ordovician time, and most of the structures of the Hudson and Lake Champlain valleys and of the ranges along its eastern margin are Taconic. The Catskills and Adirondacks, however, are part of the stable interior. See index map, Fig. 11.1 and geologic map, Fig. 11.2.

Catskill Mountains

The Catskill Mountains are west of the Hudson River and about 100 miles north of the city of New York. See geomorphic diagrams of Figs. 11.3 and 11.4. They are a dissected plateau with highest summit levels about 5000 feet above sea level and local relief of over 3000 feet. They were the site of pioneer geologic studies in North America, and in them the stratigraphic sequence of the Silurian and Devonian systems was early established. The Catskills proper consist of nearly flat-lying beds, gently inclined toward the west, and as such are part of the Appalachian Plateaus geomorphic province. The most widespread rocks are the Devonian. Along the east margin and in the adjacent Hudson Valley, the strata, especially the Cambrian and Ordovician, are highly deformed; and the Devonian and Silurian beds rest on their beveled edges. The classic angular unconformity between the Ordovician and Silurian beds, which here marks the Taconic orogeny, is displayed along the southeast margin of the Catskills. See the *Geological Map of the United States.* Also, the system of folded and thrust-faulted Appalachians of the south narrows here into a belt a few miles wide, and some of its late Paleozoic structures may here be impressed on the strata and in part superposed

Fig. 11.1. Principal physical features of New England and the Maritime Provinces. M. H. means Montarigian Hills.

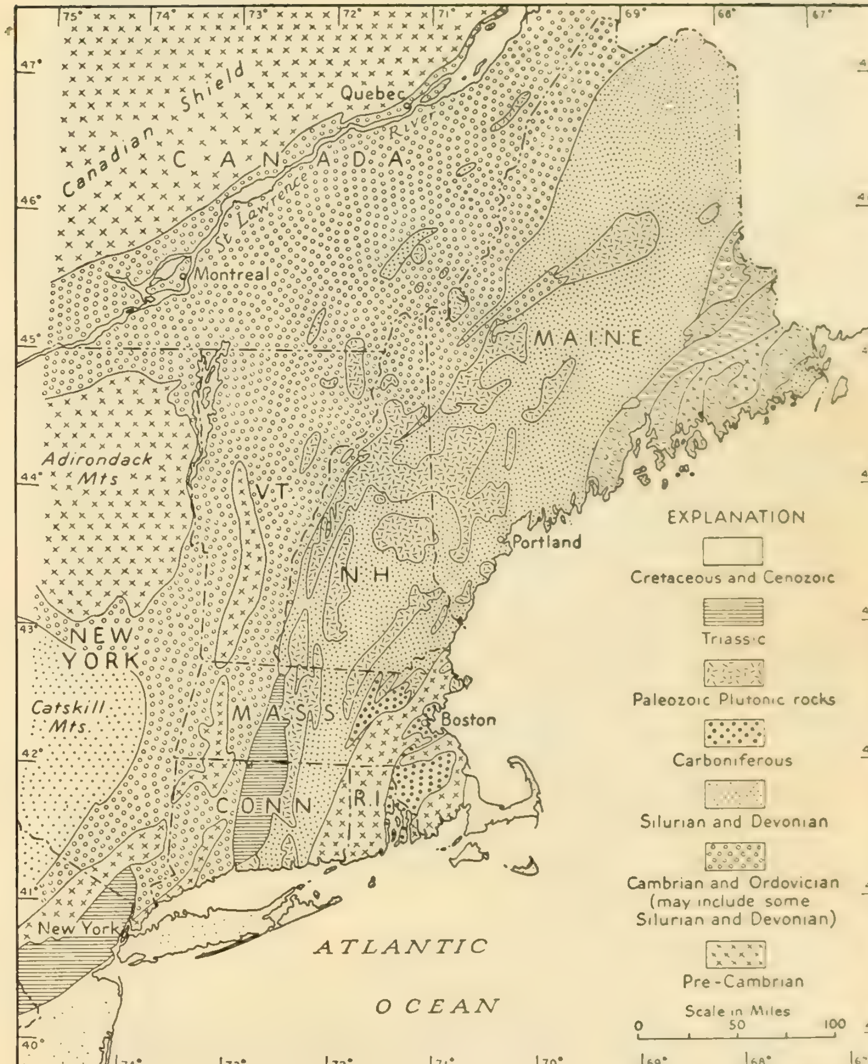


Fig. 11.2. Generalized geologic map of New England. Reproduced from Billings, 1956.

Fig. 11.3. Block diagram of lower Hudson River region by Raisz. Reproduced from *Internat. Geol. Congr. Guidebook 1*, 1933, *Eastern New York and Western New England.*

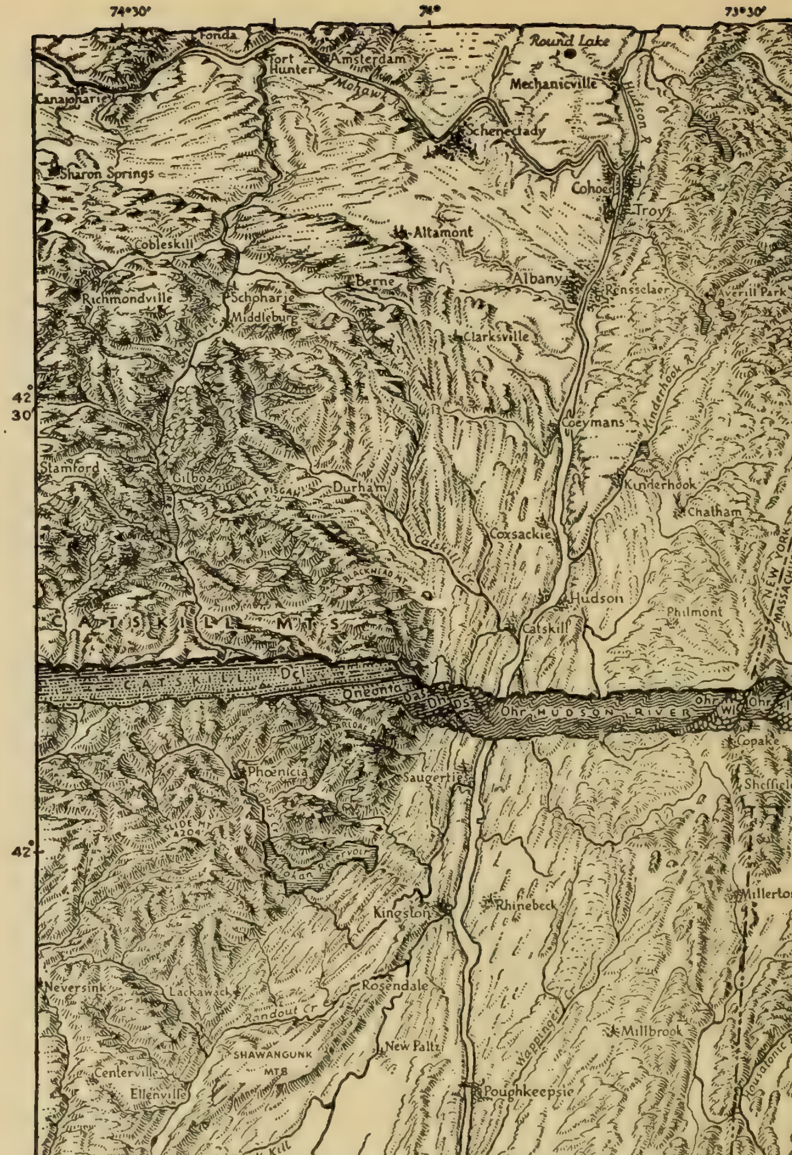


Fig. 11.4. Block diagram of lower Hudson River region. Joins opposite figure on north.

on the older structures of the Taconic orogeny. The section along the Catskill aqueduct, Fig. 11.5, gives a good idea of the composition and structure of the Catskills and adjacent Hudson Valley.

The regional stratigraphy including the Catskills has been presented in Chapter 8 on the southern and central Appalachians. See Figs. 8.10 to 8.12.

Regarding the structural history, Chadwick and Kay (1933) say the following:

> There is evidence in the region of at least two periods of deformation. In several exposures, Ordovician beds lie in close contact with angular unconformity beneath the basal Silurian sediments. Formations as young as Middle Devonian have been folded and affected by faults of low angle showing relative overthrust from the east.
>
> The first of these deformations is definitely assigned to the Taconian disturbance, for which this is the classical area of study. The later deformation may have been produced either in the Acadian disturbance at the end of the Devonian or in the Appalachian revolution, or in both. Inasmuch as late Paleozoic rocks are not present in the disturbed areas, it is not possible to date the movements precisely. The tectonic movements that produced the coarse clastic Upper Devonian sediments to the west may have been accompanied by this folding and faulting; if so, the structures are Acadian. On the other hand, the structures are similar to those formed farther to the southwest and northeast in the Appalachian revolution, and it is probable that some of the effects were produced at that time.
>
> Erosion has been dominant in the region since the end of Paleozoic time. Remnants of a peneplain may be preserved in the accordant summits of the higher peaks in the western part of the region, of which Plateau Mountain is typical. The high areas that bear these remnants seem to stand above an erosion level represented by the open upper valleys of the Catskills and by the beveled surface of the Helderberg Plateau, to the north, seen from Windham Notch. This lower level lies 2,500 feet (750 meters) below the supposed summit peneplain and has been correlated by some geologists with the Schooley peneplain of Pennsylvania, by others with the Harrisburg peneplain, of later Tertiary age. Further elevation and subsequent erosion produced a peneplain that bevels the weaker folded rocks in the Hudson River Valley west of the river. This later Tertiary surface is 1,500 feet (450 meters) below the last and has been called the Albany peneplain. More recent movements have elevated this surface a few hundred feet (100 meters or more) above present base-level, permitting the excavation of valleys in the weakest rock belts. Thus erosion has brought about the removal of a great mass of later Paleozoic sediments through several cycles of erosion with intervening uplifts, exposing early Paleozoic rocks in the eastern part of the region.

Adirondack Mountains

The Adirondack Mountains constitute a nearly circular uplift about 150 miles across, which extends from Lake Ontario on the west to Lake Champlain on the east, and from the Mohawk Valley on the south to the St. Lawrence lowland on the north. The northwestern part of the Adirondacks is a rolling upland of gentle relief and a mean altitude of about 1000 feet above sea level, whereas the southeastern part is a rugged mountain mass, individual ridges of which reach 3000 above the valley floors, and the highest peak, Mount Marcy, stands 5344 feet above the sea.

The Adirondacks consist mainly of Precambrian rocks. These are surrounded by gently upturned Cambro–Ordovician sediments, except near Kingston, Ontario, along the St. Lawrence, where a neck of the Precambrian rocks connects with the Precambrian of the Canadian Shield (the Frontenac axis) and along Lake Champlain where highly deformed strata of the Taconic system bound the dome.

According to Balk (Longwell, 1933):

> The unconformity between pre-Cambrian and Paleozoic rocks is exposed in numerous places, although in the southeast the primary relations are somewhat blurred by post-Ordovician faults along which the Adirondacks have been elevated with reference to the surrounding younger rocks. One of these faults passes through Saratoga; another one forms the escarpment northwest of town and is followed by the road from Saratoga to Glens Falls for many miles. Escarpments near Lakes George and Champlain are due to additional border faults along the eastern margin of the Adirondacks.
>
> The pre-Cambrian sedimentary rocks of the Adirondacks appear to be identical with rocks of the same general age in the Provinces of Quebec and Ontario, so that the whole region is to be regarded as an outlier of the Canadian shield.

Sedimentation in and around the Adirondack region in Cambrian and Ordovician time is illustrated in the paleographic maps of Fig. 11.6 and by the cross section of Fig. 11.7. The Adirondack dome persisted with some irregularities as an area of gentle uplift during the Cambrian and Ordovician, and by late Cincinnatian time a broad domal structure was in existence. Then the Taconic orogeny occurred along the east side and following the orogeny closely the dome was broken by block faults. Figure 11.13 is a cross section that restores the Adirondack uplift and adjacent

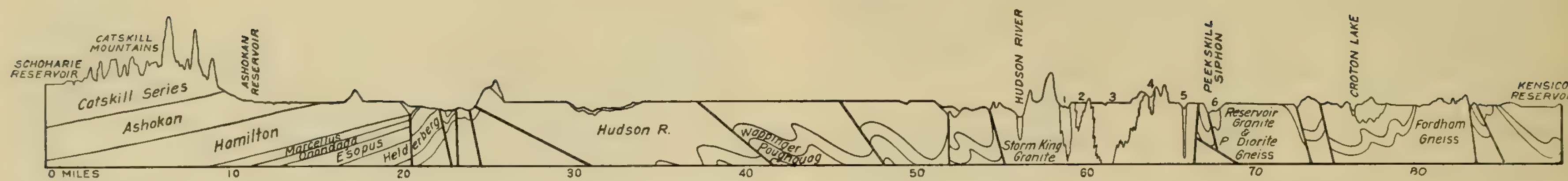


Fig. 11.5. Cross section along the Catskill aqueduct. Reproduced from *Geological Society of America Guidebook of Excursions,* 1948.

areas to this time. The distribution of faults and the Taconic front are shown in Fig. 11.12 in relation to the Lower Ordovician facies.

Lower Hudson Valley Crystallines

Definition. The block diagrams of Figs. 7.3 and 11.3 show the lower Hudson Valley area to be made up of the Triassic basin sediments and sills, and the New England upland. The following paragraphs concern the New England upland thus designated, but the term is general for much of New England, and more specific names have been given to the features of the lower Hudson Valley area. The Reading prong of Pennsylvania and the New Jersey highland merge on the northeast with the Hudson highland, whose upland surface is about 1000 feet above sea level. The Hudson River cuts a fairly narrow valley without flood plain through the highland between Newburgh and Peekskill. See map of Fig. 11.3. The Hudson highland continues northeastward into Connecticut as the Housatonic highland.

Lower Hudson Valley. From Peekskill to Manhattan Island, the Hudson is bounded on the west by the Triassic rocks, mostly thick diabase sills that form the Palisades of the Hudson, and on the east by rounded hills of a metamorphic and plutonic complex. The rocks along the route from New York City to Peekskill consist of gneisses intruded and injected by granite with infolded belts of limestone and schist. See cross section of Fig. 11.8. The major structural axes trend north-northeast and are strongly reflected in the general arrangement of ridges and valleys. Along the lower part of the river in the vicinity of Yonkers, the structures trend about N. 20° E. and are parallel with the river, but a few miles above Yonkers they strike more easterly, whereas the course of the river is nearly due north.

Hudson and Housatonic Highlands. Balk (1937) and Barth (1937) have made a thorough study of the Hudson and Housatonic highlands and adjacent areas, and report a complex of Precambrian crystalline rocks and a series of three sedimentary formations of Cambrian and Ordovician age. The highlands themselves are formed of a complex of gneisses of granitic and syenitic composition. Associated are injection gneisses as well as narrow tracts of amphibolite, marble, and other highly metamorphic rocks. Along the northwestern border of the highlands, medium- to coarse-grained granites and granite gneisses are fairly abundant.

The Paleozoic strata are described by Balk (1937) as follows:

> The oldest Paleozoic rock is a pink or white quartzite (Poughquag quartzite) that rests unconformably upon the various pre-Cambrian rocks. At the base, a conglomerate may be present, though rarely more than a few feet thick. Quartz pebbles, about an inch across, and an occasional black chert fragment, are the most abundant constituents. Fossils of Lower Cambrian age have been described from several localities in southeastern New York.
>
> The quartzite is succeeded by a sequence of carbonate rocks to which, in the Poughkeepsie area, the name, Wappinger terrane, has been applied. As elsewhere in the Appalachian region, the rocks include members of Cambrian and Ordovician age, but Quaternary deposits obscure so much of the bedrock that no complete section is available. Fossils ranging from Lower Cambrian to Middle Ordovician have been reported from various localities, but it is believed that there are several disconformities within the terrane. The thickness of the series is difficult to estimate, but may well exceed 1,000 feet.
>
> A series of slates and similar rocks, resting on the carbonate rocks, is called the Hudson River pelite. Fossils of Middle Ordovician age have been found in

the western portion of Poughkeepsie quadrangle, but farther east, cleavage seems to have destroyed them. Hudson River slates of black, gray, greenish, and red color are known; commonly, argillaceous layers are interbedded with thousands of thin, fine-grained sandy layers, or aphanitic cherty beds that weather whitish. Scattered through the series are hundreds of lenses of sandstone, or quartzite, conglomerate, and graywacke, and quartz veins penetrate the rock in almost every outcrop. On account of the intricate folding, and absence of continuous exposures, the thickness of the Hudson River series is unknown, but it may exceed that of the carbonate rocks below.

Balk's interpretation of the structure of the region may best be understood by the study of the lower cross section of Fig. 11.9. Of first importance is the unconformity at the base of the Poughquag quartzite which clearly reveals the Precambrian age of the gneiss and granite complex of the Hudson and Housatonic highlands.

The highlands are regarded as uplifted blocks. As the uplift occurred, the Paleozoic succession along the west side was tilted westward, and in addition was broken by a number of faults, most of which are thrusts of medium to steep southeasterly dip. Thrust faults are also recognized along the east flank of the northeast end of the Hudson highland. That the Precambrian highlands are uplifted masses is shown by the general basin distribution of the youngest rocks, the Hudson River pelites, in the middle of the intervening areas, and then the next older rocks, the Wappinger limestone and Poughquag quartzite next to the gneiss.

Between the Hudson and Housatonic highlands is a Paleozoic area which is regarded as a faulted syncline. It has the special significance of affording a connection between the known Cambrian and Ordovician strata on the west of the highlands to unfossiliferous and more metamorphosed strata on the east, and it is here that Balk and Barth have demonstrated the progressive metamorphism of the Hudson River slates and phyllites to schist and even injection gneisses, and the increase in marmorization of the carbonates.

The general basin structure of the strata between the masses of Precambrian gneisses is greatly marred and distorted by normal and thrust faults which have cut the quartzite for miles along the gneiss borders, and at many places have brought the limestone to the level of the pelite. Most of the faults strike north-northeast or north-south; hence, the rock

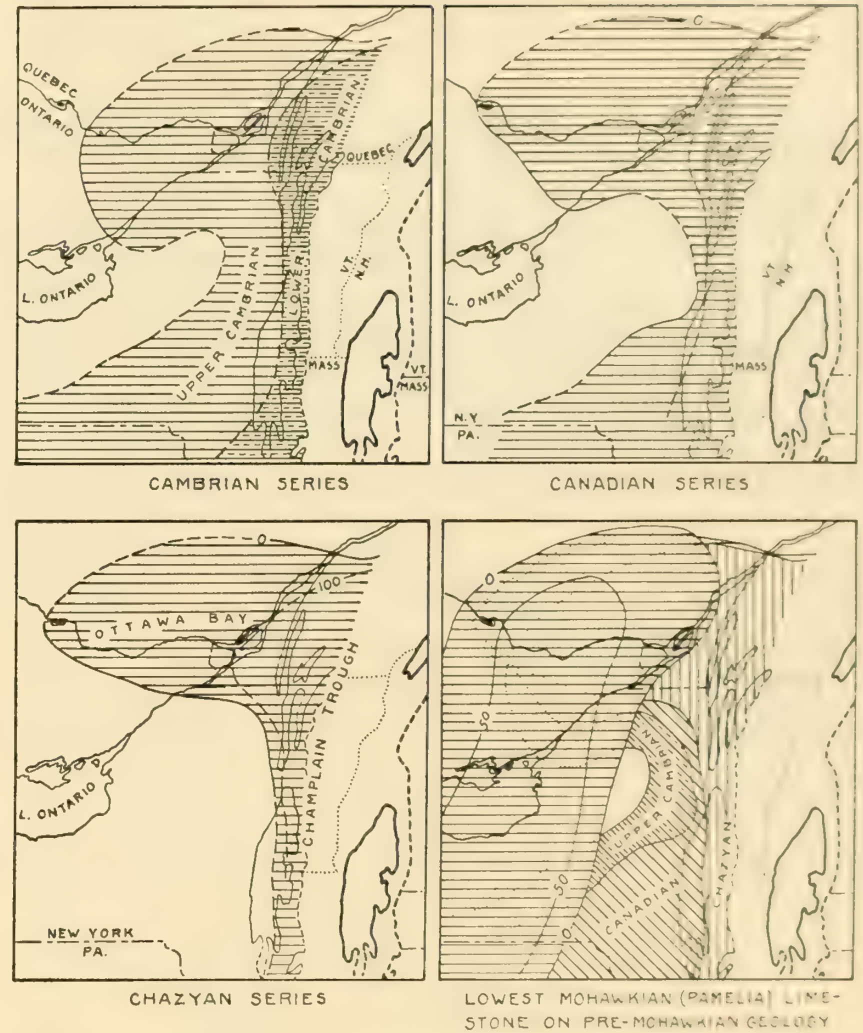


Fig. 11.6. Cambrian and Ordovician paleogeography of the New York and St. Lawrence region, after Kay, 1942. The ruled areas represent the spread of deposits, and the Taconic allochthone as postulated in Figs. 11.12 and 11.13 is shown in both present (left) and original (right) position.

units are arranged in belts of north-southerly trend. The horizontal forces that caused the thrusts are also believed to have cast the sedimentary rocks into folds which are overturned to the west. The folds, however, are very small ones in otherwise gently downfolded beds.

Cleavage pervades the crenulated sediments widely. It is everywhere parallel to the axial planes of the crenulations, and is best developed in the slate phases northeast of Poughkeepsie.

The metamorphic rocks of the lower Hudson River Valley have been regarded as Precambrian, but in light of Balk's and Barth's work it seems probable that only the Fordham gneisses is Precambrian and that the Inwood limestone is equivalent to the Wappinger limestone and the Manhattan schist to the Hudson River pelite, both of Cambro–Ordovician age. Refer to cross sections of Figs. 11.5 and 11.8. For discussion of the problem see Balk, 1937. Paige (1956) has correlated undoubted Cambro–Ordovician rocks west of the Hudson River near Peekskill with the Inwood marble and Manhattan schist east of the river.

Potassium-argon age determinations on the micas of the Manhattan schist, the Inwood marble, the Fordham gneiss, some discordant pegmatites, and a diorite were made by Long and Kulp (1958). An average age for the generation of the micas of the post-Fordham gneiss formations is given as 366 ± 9 m.y., which they say may tentatively be correlated with the Late Ordovician Taconic orogeny. Very recent interpretations by Hurley *et al.* (1959) indicate that this absolute age may be post-Early Devonian, and in connection with orogeny in New Hampshire their work will be referred to again.

Biotite from the Fordham gneiss is slightly older; the "apparent age"

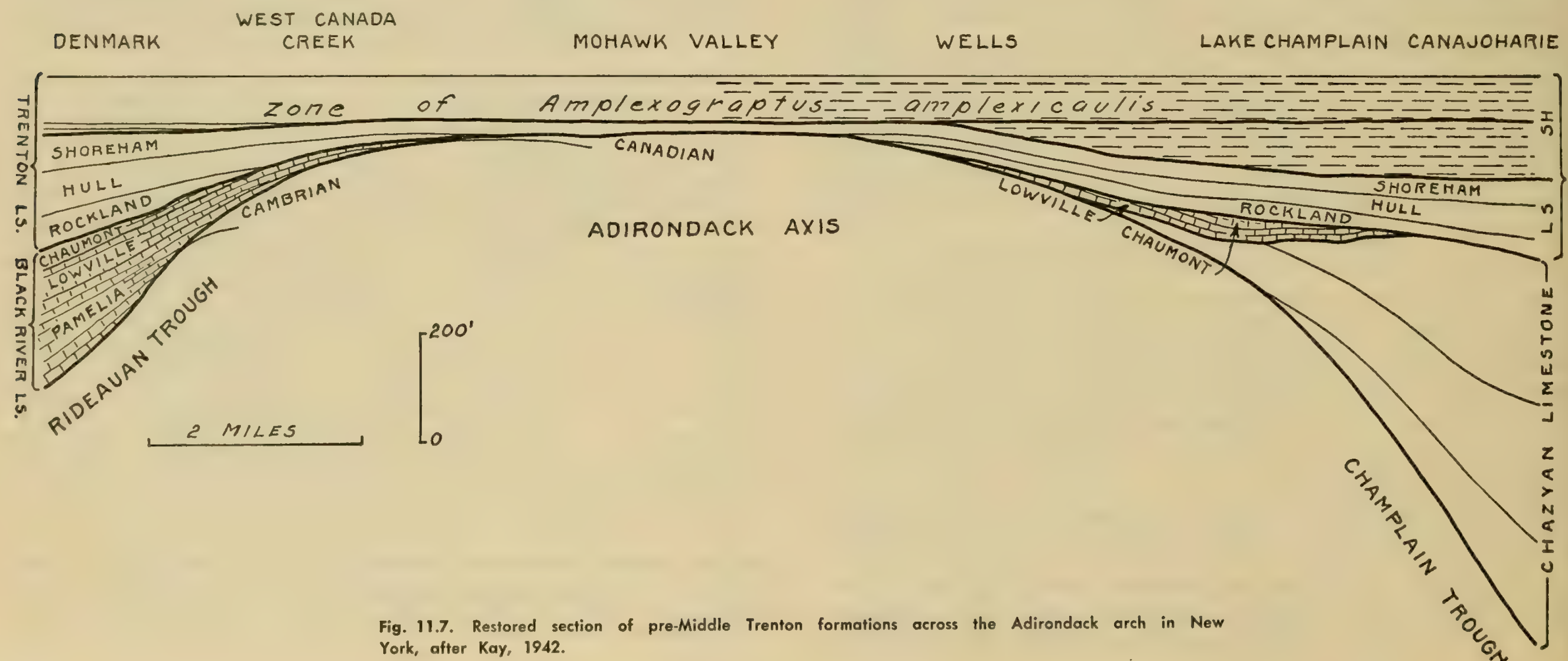


Fig. 11.7. Restored section of pre-Middle Trenton formations across the Adirondack arch in New York, after Kay, 1942.

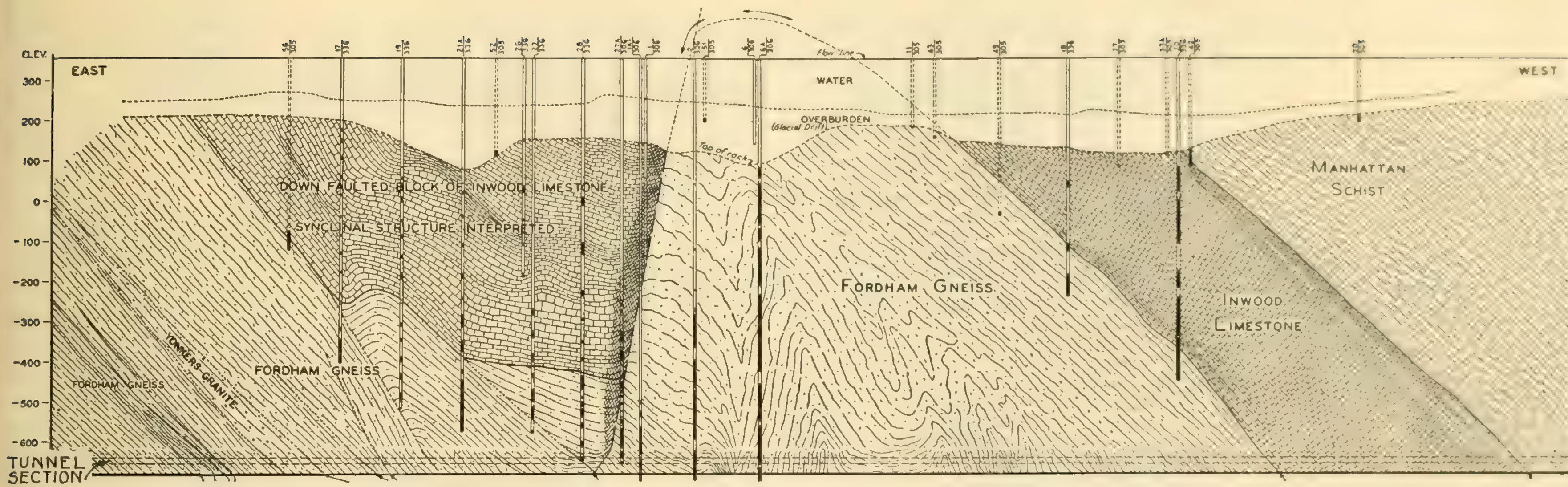


Fig. 11.8. Cross section along Kenisco bypass tunnel of the Delaware aqueduct. Kenisco Dam is just east of Croton Lake in the lower Hudson Valley. Reproduced from *Geological Society of America Guidebook of Excursions,* 1948.

of two samples is given as 400 and 440 m.y. The authors suggest that the Fordham gneiss being demonstrably older and probably the Precambrian basement did not lose all its argon during the 365 m.y. recrystallization process, and hence its micas yield somewhat older dates. It will be recalled that zircons from the Baltimore gneiss of the crystalline Piedmont yielded ages of about 1100 m.y., whereas the micas from the same rock gave ages of 300 to 350 m.y.

The age of the sediments themselves is not indicated by the isotope age determinations but, at least, the time of the last major orogeny and metamorphism is sufficiently young so that the sediments could well be Cambro–Ordovician.

Green Mountains

The Hudson and Housatonic highlands, if followed northerly, lead to the Taconic Mountains and northeasterly to the "western highland" of Connecticut and Massachusetts, of which the Berkshires are a part. See Precambrian area in western Massachusetts, Figs. 11.2 and 11.9. East of the western highland is the Triassic lowland. The Berkshire Mountains extend to the Green Mountains at about the Massachusetts and Vermont border, and the Green Mountains continue northward through central Vermont to Quebec. See Cady, 1960. The Taconic Range extends northerly along the New York–Vermont border to about central western Vermont, and between it and the Berksire–Green Mountain element is the "marble belt." For the broad relations of these geologic units see the tectonic map, Fig. 11.10. The Green Mountains are comparable in elevation with the Adirondacks which lie across the Lake Champlain Valley, but the other highlands and ranges are comparatively low.

The core of the southern end of the Green Mountains is made up of granites and gneisses of Precambrian age. These ancient rocks are overlapped on the flanks by quartizites of lowest Cambrian age. See lower cross section of Fig. 11.11. The northern part of the range is a gneiss and schist anticlinorium which plunges northerly, and although somewhat like

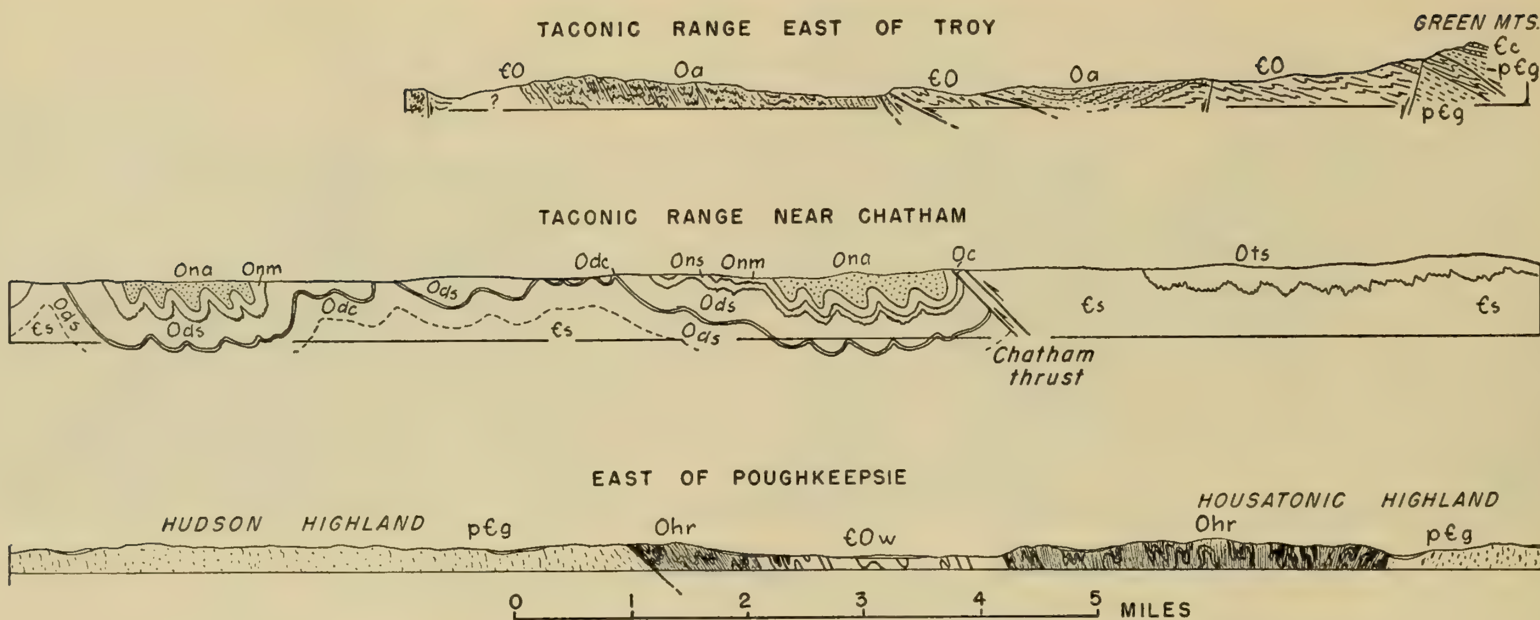


Fig. 11.9. Cross sections of central and southern Taconic Range. Section east of Troy, N. Y., after Balk, 1953. p€g, Precambrian gneiss; €c, Lower Cambrian Cheshire quartzite; €O, Cambro-Ordovician limestone and dolomite; Oa, gray, purple, and black slate and quartz-chlorite schist.
Section near Chatham, N. Y., after Craddock, 1957. €s, green slate with interbedded graywacke and quartzite; Oc, carbonate rock; Odc, green shale; Ons, red shale member; Onm, Mount Merino dark shale wtih interbedded chert; Ona, Austin Glen graywacke and dark shale.
Section east of Poughkeepsie, N. Y., after Balk, 1937; p€g, Precambrian gneiss; €Ow, Wappinger dolomitic limestone; Ohr, Hudson River pellite, phyllite, and schist.

the southern core is believed by Cady (1945) to be part of the Taconic allochthone. See map of Fig. 11.10. In its east flank the Green Mountain anticlinorium contains a discontinuous belt of ultrabasic intrusives which are associated with volcanics including pillow basalt.

Taconic Mountains

The Taconic Mountains are a low range of hills composed mostly of argillaceous rocks such as phyllite, slate, and shale. This clastic sequence is surrounded in the adjacent lowlands by rocks, chiefly carbonates. In the Taconic sequence, as it is called, there is one thin quartzite formation and one very thin limestone which together form perhaps 5 percent of the section. There are three slate formations of Middle Ordovician age and six of Lower Cambrian. No Middle or Upper Cambrian is present and no Lower Ordovician. The Lower Cambrian of the Taconic Range lies beside the Lower Cambrian of the valleys and the two groups have no features in common except that of age (Keith, in Longwell, 1933). Similarly, most of the Ordovician of the mountains differs from the Ordovician of the surrounding valleys. These relations have led through a long controversy to the interpretation of the Taconic clastic sequence as a klippe, which represents an eastern trough facies that has been thrust westward 30 to 50 miles or more on a western trough sequence. It is part of the Taconic allochthone. The carbonates of the western trough supposedly are the autochthone. See cross section D–D′, Fig. 11.11. The details and relations will be taken up later.

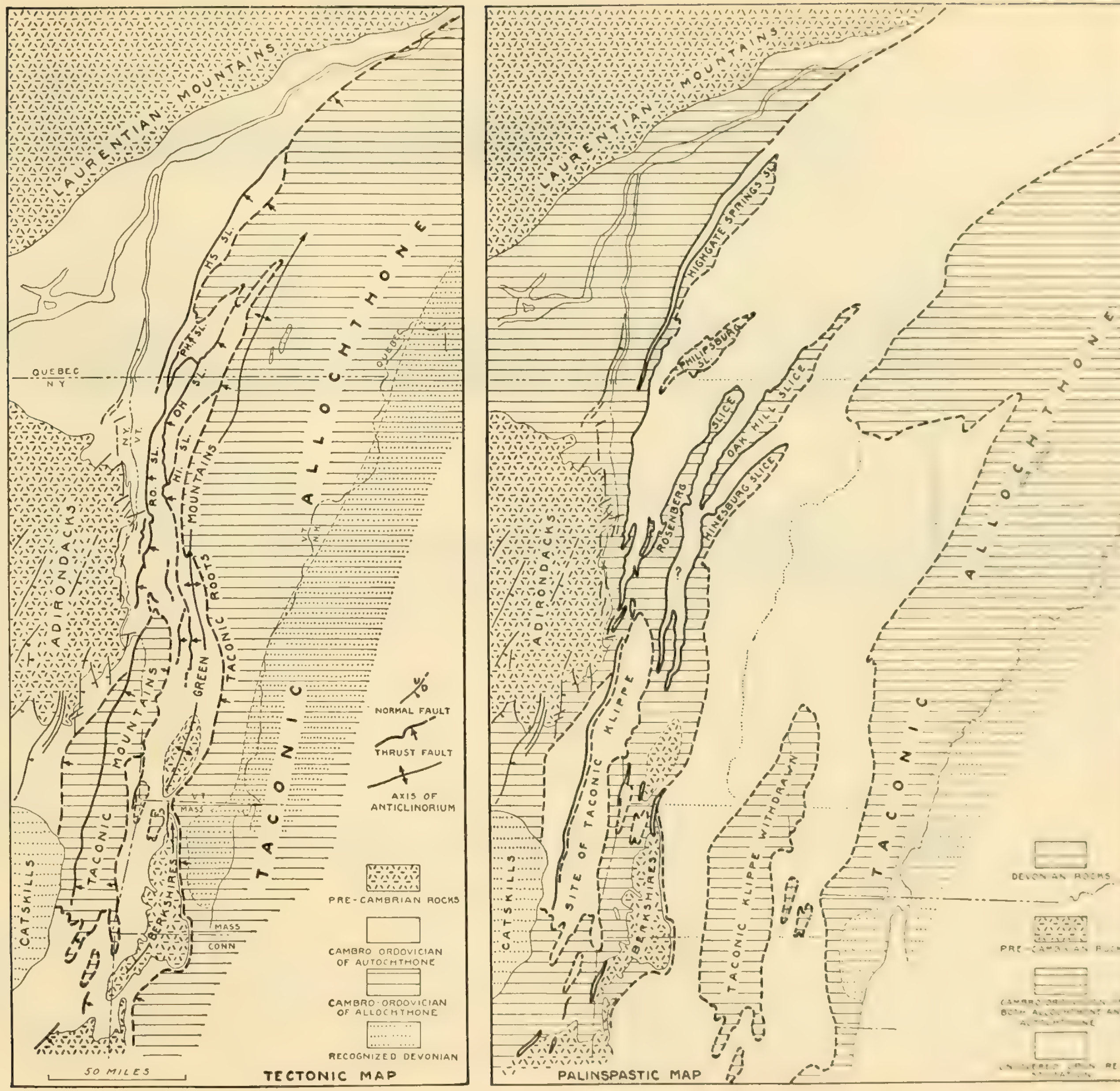


Fig. 11.10. Tectonic and palinspastic maps of the Taconic system in eastern New York, western New England, and southern Quebec, after Cady, 1945. The palinspastic map attempts to restore the thrust slices to their approximate position before they were moved westward. Since the Devonian strata were deposited after the Taconic orogeny, they were not displaced by it and do not participate in the restoration. S.L. means slice or thrust sheet and the abbreviations in the tectonic map may be identified by comparison with the palinspastic map.

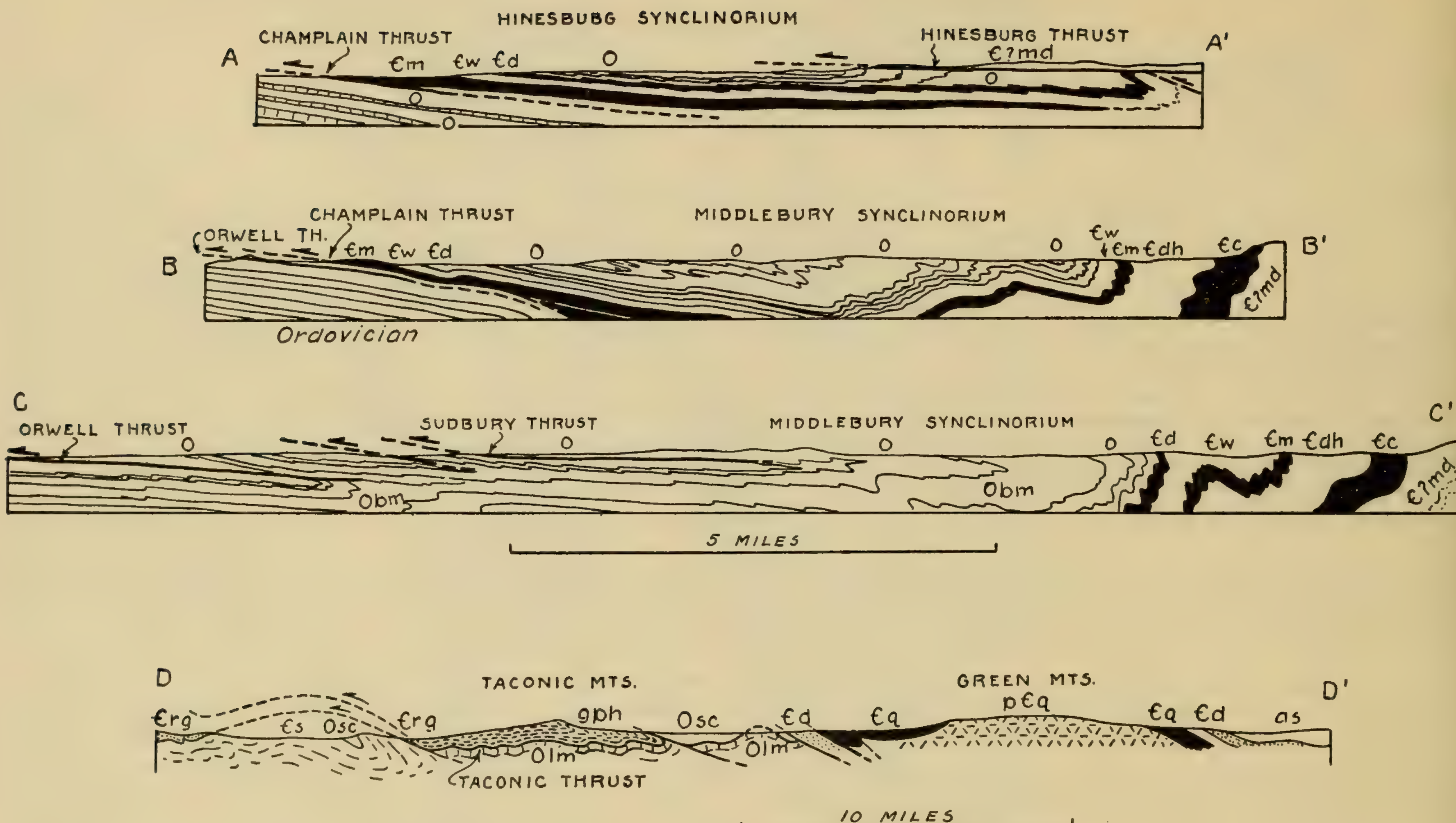


Fig. 11.11. Cross sections of the Taconic system of western Vermont, A–A′, B–B′, and C–C′, after Cady, 1945. Refer to map of Fig. 11.16. Ꞓ?md, Mendon series; Ꞓc, Cheshire quartzite; Ꞓdh, Dunham dolomite; Ꞓp, Perker shale; Ꞓm, Monkton quartzite; Ꞓw, Winooski dolomite; Ꞓd, Danby formation; O, several Ordovician formations; Obm, Bascom formation.

Cross section of Taconic and Green Mountains along Vermont-Massachusetts border and into eastern New York, D–D′, after Knopf and Prindle in Longwell, 1923. pꞒq, granite gneiss; Ꞓq, quartzite, including phyllite and conglomerate; Ꞓd, dolomite; Ꞓrg, graywacke; Ꞓs, black shale; Olm, limestone and marble; Osc, black shale, red shale, and chert; gph, Cambrian (?) green phyllite; as, Cambrian (?) albite schist.

The manner of thrusting, as conceived by Kay, in map view is graphically illustrated in Fig. 11.12, and in cross section in Fig. 11.13.

Two cross sections of the central and southern Taconic Range are presented in Fig. 11.9 and should be referred to in the following discussion against the klippe hypothesis.

In a study of the Taconic Range west of Troy, Balk (1953) recognizes thrusting and an eastern allochthonous sequence and a western autochthonous sequence, but concludes that dense vegetation cover and much drift leave so few outcrops that the existence of a great Taconic klippe cannot be proved or disproved. In a study farther south near Poughkeepsie (1937) he believes there is little evidence to support the thrust and klippe hypothesis.

> Thrust sheets and klippen are postulated because of anomalous stratigraphic successions, not otherwise explainable; or because of structure anomalies not understandable from other points of view; or because the klippen, although closely related to rocks nearer the root zones, were obviously out of their proper geologic setting; or on the evidence of intensely crushed subhorizontal zones of deformation; or on the evidence of exposed soles. None of these criteria appears to be fully applicable here. There is no proof of an anomalous stratigraphic succession in the gap of Wingdale; the deformation of the supposed thrust sheet of pelite is, to all appearances, synchronous with that of the autochthonous formations; the gneiss of the supposed klippen is known to underlie the sedimentary rocks a few thousand feet below the surface; no crush horizons, or exposures of indubitable soles, have been observed (Balk, 1937).

Craddock (1957) also concludes against the major klippe hypothesis (middle cross section of Fig. 11.9) in a study of the southern end of the Taconic Range. He says:

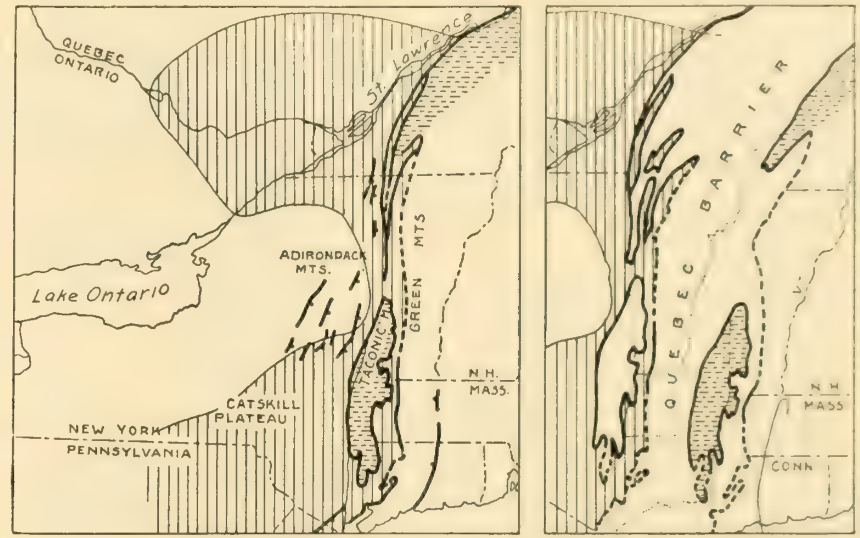


Fig. 11.12. Distribution of Canadian (Lower Ordovician) facies in New England and New York, after Kay, 1942. The map on left shows the present distribution as a result of the Taconic (post-Ordovician) thrusting, and the map on right shows the inferred distribution before thrusting (a palinspastic map). Vertically ruled sediments are carbonates; horizontally dashed sediments are shales.

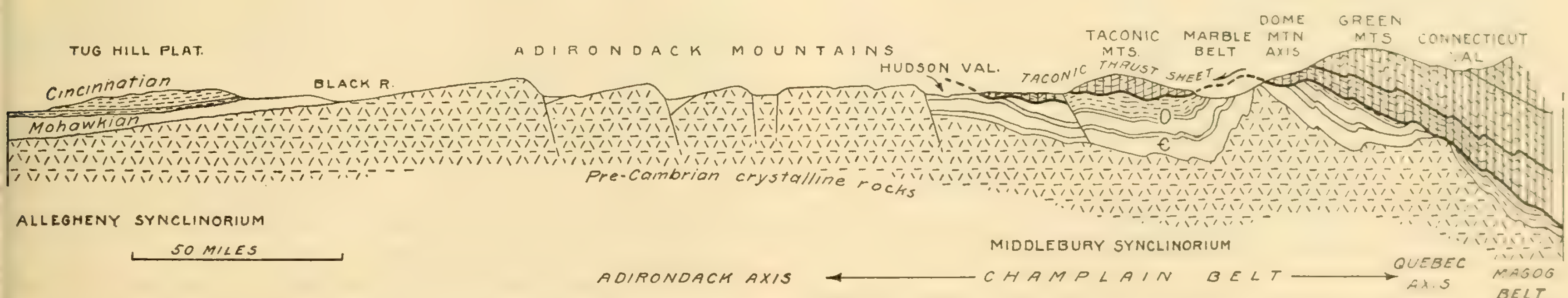


Fig. 11.13. Section of Adirondack dome and Taconic system restored to early Silurian time (after Kay, 1942).

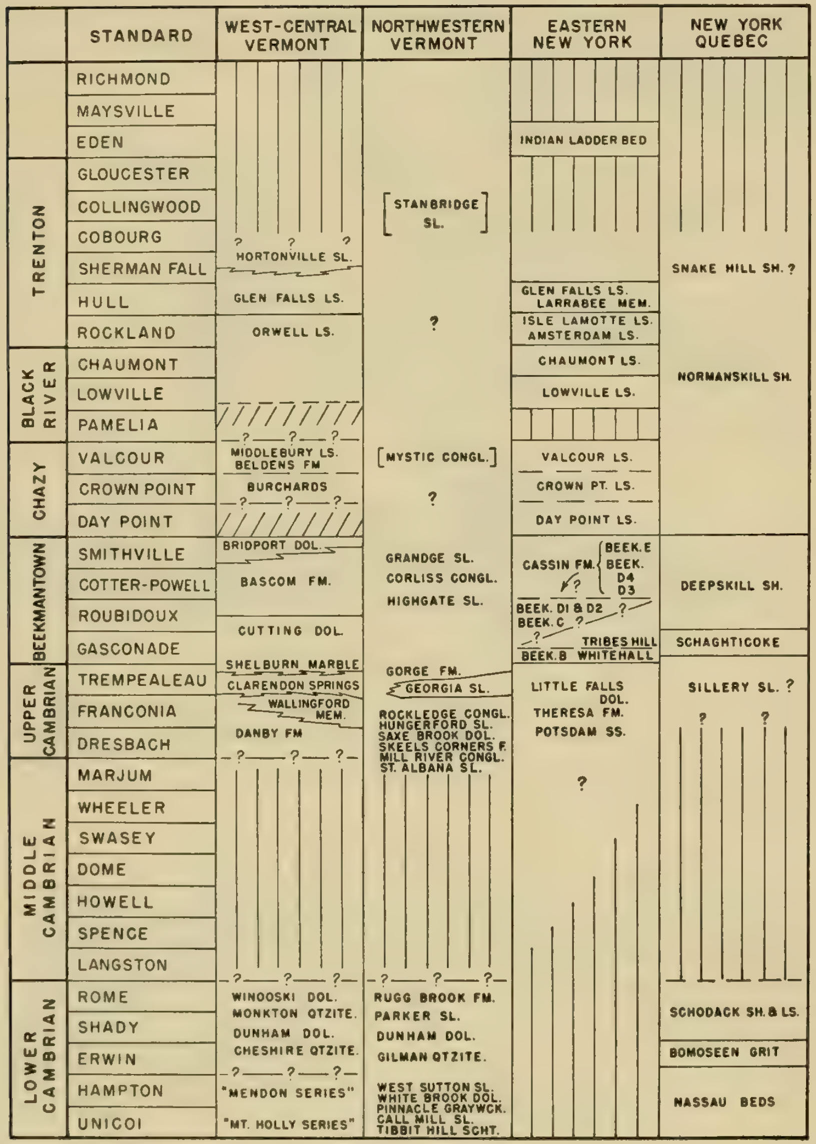


Fig. 11.14. Stratigraphic correlations in west-central Vermont and adjoining areas, after Cady, 1945.

Evidence for the existence of the "Taconic klippe" was not found in mapping this quadrangle. Analysis of the development of the klippe hypothesis indicates that it is based principally upon stratigraphic considerations; available structural evidence weighs against this interpretation. While the klippe hypothesis seems to explain well the relations at the north end of the Taconic Range, the problem of adequately defining the boundaries of this "klippe" causes serious doubt about its existence.

An alternative interpretation of the regional relations is suggested, involving unconformities and facies changes in a single indigenous sequence. Trentonian rocks lie unconformably on rocks as old as Precambrian from Vermont to Pennsylvania and pass indiscriminately in and out of the "Taconic klippe." The Normanskill and Deepkill rocks (mainly shale) are interpreted as passing transitionally into limestone to the west. The Deepkill is believed to rest unconformably on rocks of Early Cambrian to middle Canadian age. Middle Canadian formations in the kinderhook quadrangle are carbonate rocks and appear to rest unconformably upon Lower Cambrian slates; their striking similarity to equivalent rocks in the near-by "autochthonous" series suggests they have not been displaced any great distance. The lower Cambrian is a thick series of argillite, graywacke, and quartzite with some thin carbonate rocks near the top. The thick, lower part of this series is considered a southward continuation of the Mendon Series of Vermont. The upper strata are interpreted as the offshore equivalents of shallow-water quartzites and carbonate rocks deposited marginal to an eastern welt in later Early Cambrian time.

Lake Champlain and St. Lawrence Lowlands

The Champlain Valley lies partly in New England. In the largest view it is bounded on the east by the Green Mountains and on the west by the Adirondack Mountains, and at the south it is split by a minor group of mountains, the Taconic Range. A large part of the valley is occupied by Lake Champlain, the surface of which is 100 feet (30 meters) above sea level and the bottom is below sea level. The valley passes northward into Canada and curves northeastward, merging into the St. Lawrence Valley (Keith, in Longwell, 1933).

The valley is divided by the Taconic Mountains into a western part which is continuous with the Hudson River Valley, and an eastern part which extends along the eastern side of the range nearly to Long Island Sound. This eastern part of the valley is known as the Rutland Valley in Vermont and the Stockbridge Valley in Massachusetts.

The St. Lawrence lowlands are of two divisions separated by the fault known as Logan's line (Fig. 12.2). Southeast of the fault is the deformed

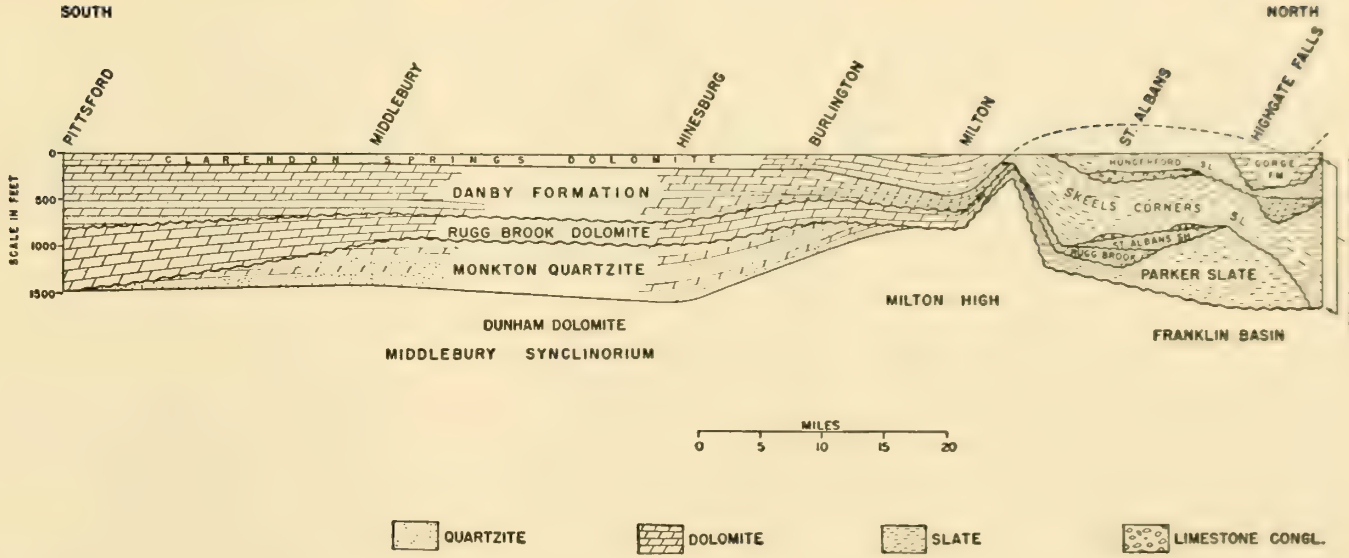


Fig. 11.15. North-south section in northwestern Vermont of the Cambrian and Lower Ordovician formations, restored to early Ordovician. Reproduced from Shaw, 1958.

Taconic belt and northwest of it is the undeformed shelt sediments which lay onto the Canadian Shield.

Stratigraphy. The stratigraphic columns presented in Fig. 11.14 are by Cady (1945) and represent a long endeavor by numerous geologists to unravel the succession and to correlate the different formations in the region. As previously noted, it appears that two lower Paleozoic successions of approximately equivalent age exist within the same area, and the tendency of most workers is to regard the argillaceous sequence as an allochthone from the east now reposing on a calcareous western sequence. The Cambro–Ordovician limestones and dolomites grade westward into foreland sandstones of the Adirondack area, and, it is believed, eastward into shales of geosynclinal thickness. Cambrian and early Ordovician sandstone tongues extend far to the east. The geosynclinal trough migrated westward later in Ordovician time and resulted in the deposition of a shale facies over, and in places uncomformably on, the calcareous and sandy succession. This was the occasion of the Vermontian disturbance (Kay, 1942).

Kay's (1942) map of Fig. 11.11 restores the distribution of Lower Ordovician strata in the region. He names the eastern trough in which the shale facies was deposited, the Magog; a postulated barrier to the west, the Quebec; and the shallower trough in which the carbonates were deposited, the Champlain.

In the St. Alban's area of northwesternmost Vermont, north of Cady's mapping, Shaw (1958) reports some unexpected facies changes in the Cambrian and Lower Ordovician along the structural strike. These are illustrated in Fig. 11.15. A northern basin, the Franklin, was partially restricted from a southern by an east-west high, the Milton, and streams carried considerable clastic material into it from the Adirondack and Laurentian land area. Throughout Cambrian and Early Ordovician times the basin was one of considerable crustal unrest, as evidenced by the several unconformities.

Structure. Although considerable doubt exists about the Taconic klippe hypothesis south of Albany, there seems little question in the minds of those who have worked in the Lake Champlain lowlands about the reality of major east to west thrusting.

A number of thrusts other than the great Taconic thrust, but of the

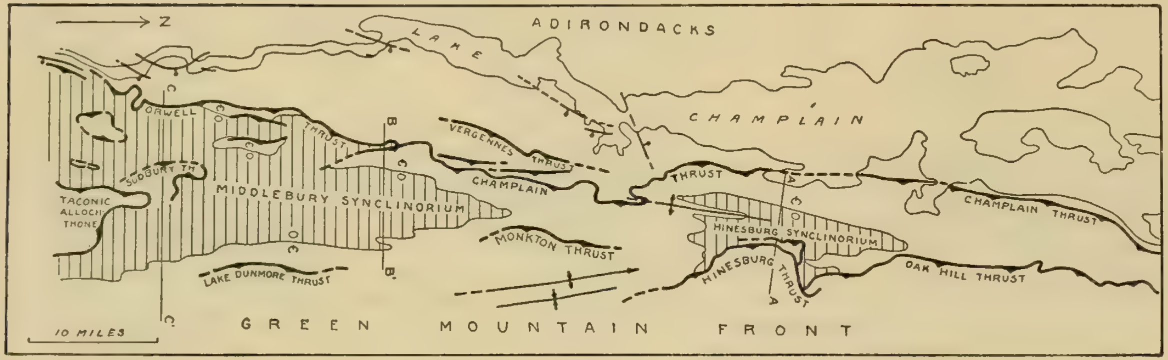


Fig. 11.16. Tectonic map of west-central Vermont, after Cady, 1945. Ruled areas are Ordovician strata in the synclinoria. Faults with knobbed bars are normal faults.

same orogeny, have been mapped. All are interpreted as having moved from east to west. The chief ones of these are the Champlain and Hinesburg–Oak Hill. The Champlain thrust trends parallel to Lake Champlain and extends from a point near the south end of the lake northward about 60 miles to the Canadian border. About 3 miles north of the border, near the village of Rosenberg, it becomes obscure in a shale terrane (Cady, 1945). See Fig. 11.16. The thrust at the north end is known as the Rosenberg slice (sheet). According to Cady, near the south end:

> At Snake Mountain Lower Cambrian beds of the mountain proper are thrust westward across and beyond Upper Cambrian and Beekmantown rocks of the Orwell thrust plate onto the Middle Trenton limestones and shales next west and structurally continuous with those found along the lake; the Champlain thrust apparently truncates the Orwell thrust.

The Hinesburg–Oak Hill thrust complex floors a tectonic unit east of the Champlain thrust and it in turn is bounded on the east by the Green Mountains. The southern part is called the Hinesburg thrust, and the northern the Oak Hill. The Oak Hill thrust sheet passes beneath the Hinesburg.

> The rocks of both the Hinesburg and Oak Hill thrust slices grade eastward into the schist and gneiss terrane of the Green Mountains. Both of these slices, so far as they have been delineated, apparently have undergone considerable displacement, as evidenced by the depth of erosional re-entrants and by the outlying position of klippes. The Hinesburg and Oak Hill thrusts form the eastern boundary of the Rosenberg slice.
>
> The rather highly deformed quartzose slates, phyllites, and graywackes east of the Hinesburg thrust, a short distance north and east of Hinesburg village, lie with angular discordance across the east limb of the Hinesburg synclinorium, where the thrust plane truncates minor folds which are made up of beds from Lower Cambrian to Beekmantown age. The thrust plane has not been observed at any point, but the depth of the re-entrants suggests that it dips at a very low angle to the east. Non-quartzose black slates and phyllites crop out west of the quartzose rocks along the thrust front in St. George and Williston townships. These latter Upper Cambrian argillaceous rocks comprise the Muddy Brook thrust slice, which was apparently dragged up along the sole of the Hinesburg thrust. These same slates and Upper Cambrian sandy dolomites crop out in the re-entrant west of Williston village. Northwest of Williston village the quartzose rocks are thrust over a closely folded syncline of the Oak Hills slice. In this syncline are formations from Lower Cambrian to probably Upper Cambrian age.
>
> In general, the rocks east of the Oak Hill thrust are less deformed and less uniform in appearance than those east of the Hinesburg thrust. The lower Cambrian Dunham dolomite is everywhere recognizable, and at many places along the thrust front, where structures involving the Dunham are truncated at erosional re-entrants or at klippes such as Cobble Hill in Milton township, it locates the fault. Where argillaceous rocks are near the contact, the fault is much more difficult to locate, inasmuch as the eastern exposures of the Rosenberg slice are in a predominantly argillaceous terrane (Cady, 1945).

Two synclinoria lie on a common north-south axis and are separated by the Monkton cross anticline. See Fig. 11.16. They are bounded on the west by the Adirondack dome and Champlain thrust and on the east by the Hinesburg–Oak Hill thrust and the Green Mountains.

The southern synclinorium, known as the Middleburg, makes up the

structure of the area between Snake Mountain on its west limb and the Green Mountain front on its east limb. The center of the synclinorium is covered by the great Taconic klippe south of the latitude of Brandon. The east limb may be traced fairly continuously into the marble belt south of this latitude (Cady, 1945). The west limb loses its identity in an area of high angle faults southwest of Orwell. The nature of the numerous small folds of the synclorium are best shown in the cross sections B and C of Fig. 11.11.

The northern synclinorium, known as the Hinesburg, composes the structure of most of the area between Lake Champlain and the Green Mountain front. See section A, Fig. 11.11. Most of the east limb is covered by the Hinesburg-Oak Hill thrust slices. The Hinesburg synclinorium is not so symmetrical as the Middleburg synclinorium, and the folding is limited to the development of a series of moderately broad basin structures (Cady, 1945).

The normal faults of the Adirondacks have already been described. The eastern border of the crystalline mass is formed in part by these faults, and they seem to be genetically related to the uplift of the dome. They do not intersect the major thrusts of the Lake Champlain region, but they parallel the Orwell and Champlain thrusts, and for a distance the bends in the normal faults coincide with bends in the thrust fronts. It is suggested (Cady, 1945) that the thrust fronts may have retreated by erosion eastward after they were trimmed by the normal faults, and thus the parallelism has resulted.

Tectonic History

Champlain and Magog Troughs. In 1923 on the occasion of his presidential address on North American geosynclines, Schuchert postulated a western trough, the St. Lawrence, through the Lake Champlain and St. Lawrence region, a medial divide or geanticline, and then an eastern trough, the Acadian, principally through Nova Scotia and New Brunswick. The geanticline included the Green Mountains of Vermont and the White Mountains of New Hampshire and Maine. The rocks of these mountains were then regarded as Precambrian. Since then several groups of fossils have been found, and most of the metamorphosed sediments of Schuchert's geanticline have turned out to be Lower and Middle Paleozoic in age. Still two troughs seem necessary, but the eastern one must have occupied approximately the site of Schuchert's geanticline. It has been called the Magog eugeosyncline by Kay (1942), and the western has been called the Champlain miogeosyncline. The Magog is characterized by shales, cherts, and various volcanics, the western by carbonates. Until Mid-Ordovician time, the separation of the two troughs was probably a matter of facies, but then a land barrier called Vermontia rose within the western part of the Magog trough and caused the deposition of clastics over the carbonates of the western trough. See Fig. 11.17. Later in

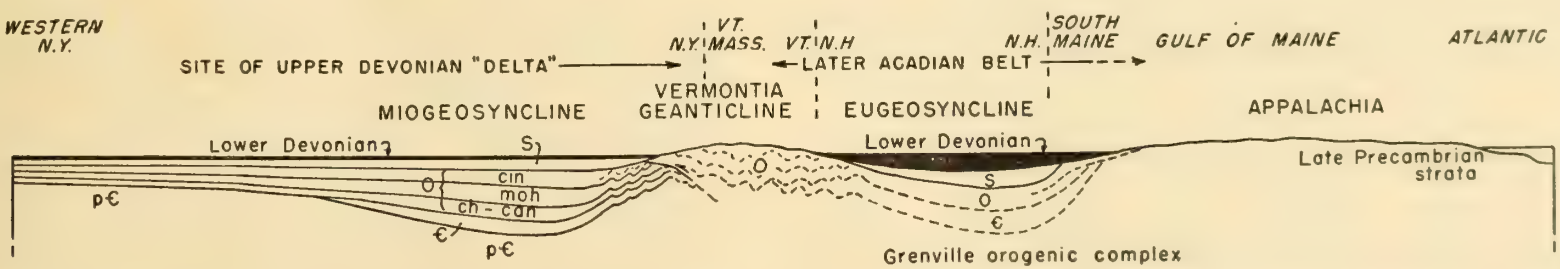


Fig. 11.17. Basins of deposition across New England just prior to Acadian orogeny. Compiled from Kay (1951), Billings (1956), and other sources. Vermontia had risen in Mid-Ordovician time and evidently was considerably wider than present dimensions indicate to supply the voluminous clastics to the miogeosyncline in Mid- and Late Ordovician time. Vermontia as shown was also essentially the site of the Taconic orogeny at the close of Ordovician time. The eugeosyncline was the site of much volcanism, and Vermontia the site of ultramafic intrusions. cin, Cincinnatian; moh, Mohawkian; and ch-can, Chazyan and Canadian. The region of Vermontia in places probably received Silurian and Devonian sediments, so its history and nature is complex.

	NORTHWESTERN VERMONT Montpelier Quad. Cady, 1956	CENTRAL AND EAST CENTRAL VERMONT White and Jahns, 1950	WESTERN, CENTRAL AND NORTHERN NEW HAMPSHIRE Billings, 1956	SOUTHEASTERN NEW HAMPSHIRE Billings, 1956
LOWER DEVONIAN		?Meeting House slate Gile Mountain fm.	Littleton fm. 15,000'±	Littleton fm. 15,000'±
SILURIAN	Waits River fm. Northfield slate	?Standing Pond vols. Waits River fm. Northfield slate	Fitch fm. 0-769'	Berwick fm. 10,000'± Eliot fm. 6,500'±
ORDOVICIAN	Shaw Mtn. fm. Serpentine, talc-carbonate rock, and steatite	Shaw Mountain fm. Ultramafic rocks	Clough quartzite 0-1200' Partridge fm. 0-2000' Ammonoosuc vols. 2000-5000'	Kittery quartzite 1,500'± Rye fm. 2,000'±
	Moretown fm. Stowe fm.	Cram Hill fm. Arenites of the Braintree-Northfield Range	Albee fm. 5000' Orfordville fm. 3500-4000'	
CAMBRIAN	Ottauquechee fm. Camels Hump gr.	Ottauquechee phyllite Pinney Hollow schist Quartzose schist, quartzite, dolomite, and conglomerate		
PRECAMBRIAN	(To the southwest)			

Fig. 11.18. Correlation chart of pre-Acadian Paleozoic formations across Vermont and New Hampshire. The Standing Pond volcanics and Meeting House slate are listed by Billings for westernmost New Hampshire in the stratigraphic order shown but not included by Cady for Vermont. The total Vermont section is immensely thick.

Ordovician time, another uplift, the Oswegan disturbance, occurred and spread westward past the Adirondack axis into the Allegheny basin.

Taconic Orogeny. At the close of the Ordovician period the major Taconic orogeny occurred, and the argillaceous rocks of the Magog trough were thrust far westward. The Quebec barrier and eastern part of the Champlain trough were concealed by it. The amount of horizontal displacement probably exceeded 40 miles (Kay, 1942).

The thrust sediments are in tectonic contact on Queenston shale in southeastern Quebec, and the autochthonous Cincinnatian has been folded considerably. The overthrust rocks are overlain at Becraft Mountain, New York, and in the Catskill Front by latest Silurian Manlius limestone. Thus, there is direct evidence that the principal lateral movements were pre-Manlius and post-Queenston. Folds in autochthonous Ordovician are truncated by the Shawangunk and Tuscarora quartzites of the earliest Silurian in southeastern New York and Pennsylvania; if the folding accompanied Taconic thrusting, the revolution is pre-Silurian.

The front of the thrust sheet is not very high. Middle Ordovician sediments are preserved near to the westernmost remnant of the sheet and probably never were buried deeply. On Anticosti Island in the Gulf of St. Lawrence, there is essentially continuous section of Cincinnatian and early Silurian calcareous shale and limestone in the Champlain belt within 50 miles of the overthrust rocks of Gaspé; the allochthone was beneath the sea or not high enough to produce significant detritus after the revolution. Though the quantity of Silurian terrigenous sediments is distinctly smaller than that of the Ordovician, . . . this reflects repeated uplift and continued presence of Vermontian highlands in later Ordovician, in contrast to progressive reduction of the transposed Taconia in the Silurian. The greatest quantity of eroded material was laid down in the latitude of Pennsylvania, as shown by isopachs; that the greatest elevation was there is also shown by the coarser texture of the sediments. The lateral movement of the allochthone may have been as great or even greater in Quebec, but Vermontia and its transposed descendant, Taconia, were more continually high farther south.

Acadian Orogeny. The next great influx of clastic sediments was in the Middle Devonian, and the sediments generally coarsen upward and eastward. They came from rising highlands on the east. The elevation terminated in the Acadian orogeny which was followed by the deposition of Mississippian clastics to the west of the orogenic belt.

The Acadian belt is known best in New Hampshire and the Maritime Provinces and will be described later, but it is possible that it spread westward to the Hudson Valley and Lake Champlain lowlands and impressed additional folds on the Taconic structures. It is possible, also, that the later structures are Appalachian in age.

Unsolved Problems. The above summary of the history of the Taconic system savors of those who postulate the great Taconic allochthone, and this is the general opinion of those who have worked in northern Massachusetts, Vermont, and eastern New York. Yet Balk and Craddock in very thorough work, at the south end of the Taconic klippe where the great thrust and its roots should be found, do not find evidence of it, and they do not believe the thrust theory necessary to explain the facies and metamorphism there. Similarly, the roots of the thrust are not yet established at all well in the Green Mountains.

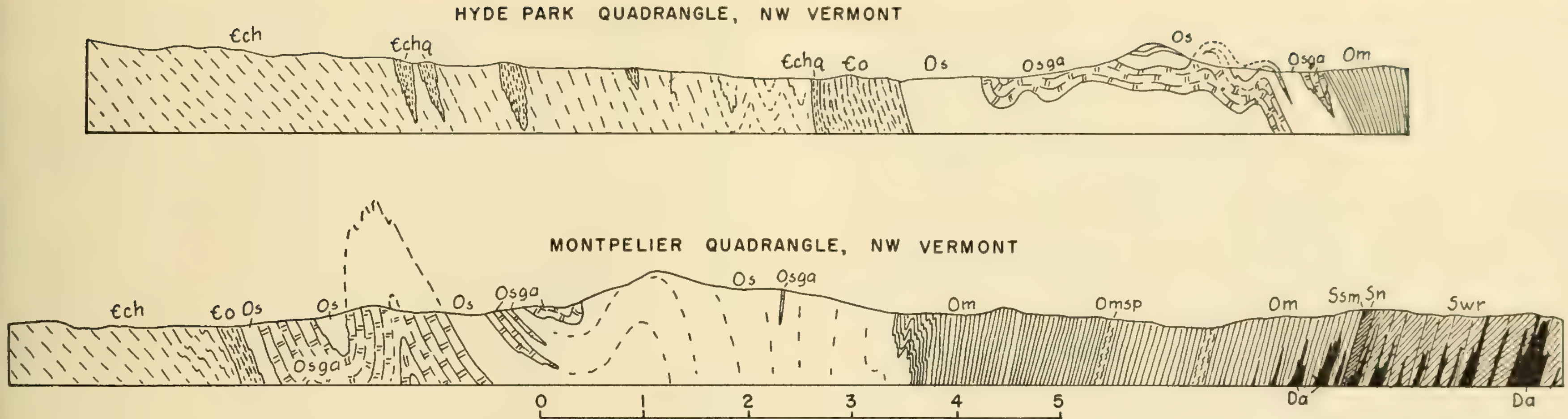


Fig. 11.19. Cross sections of northwestern Vermont in Green Mountains. Hyde Park section after Albee, 1957. Montpelier quadrangle after Cady, 1956. €ch, Camels Hump group; €chg, albite and tremolite greenstone; €o, Ottanquechee fm.; Os, Stowe fm.; Osga, middle unit of Os consisting of greenstone and amphibolite; Om, Moretown fm.; Omsp, carbonaceous and slate member; Ssm, Shaw Mountain fm.; Sn, Northfield slate; Swr, Waits River fm.; Da, Adamant granite.

CENTRAL AND EASTERN NEW ENGLAND

Definition

The Acadian orogeny of Late Devonian time affected much of New England and the Maritime Provinces, and undoubtedly spread southward through the Piedmont crystalline province of the Atlantic margin. It created a mountain system that was superposed in part on the earlier Taconic system. Where best known and perhaps best displayed in New Hampshire, New Brunswick, and Nova Scotia, it is an irregular north-south belt east of the Taconic system, but its western limit is as yet poorly defined.

The region here discussed lies east of the crest of the Green and Berkshire Mountains and includes the New England seaboard lowland, the New England upland and the White Mountains in the United States and Canada. See map of Fig. 11.1 and 11.2. The seaboard lowland extends along the Atlantic coast as a narrow zone from Rhode Island to the border of Maine and New Brunswick.

Stratigraphy and Structure of Vermont

An immensely thick section of stratified rocks exists in northwestern, central, and east-central Vermont, probably reaching a thickness of 100,000 feet (White and Jahns, 1950). The strata except some lamprophyre dikes are folded and metamorphosed sedimentary and volcanic rocks. A number of units, members or formations of volcanic rock throughout the section from Cambrian to Lower Devonian attest the eugeosynclinal nature of the deposits. See correlation chart of Fig. 11.18.

Northwestern Green Mountains

Two quadrangles, the Hyde Park and Montpelier, have been mapped by Albee (1957) and Cady (1956), and depict the structure and stratigraphy near the north end of the Green Mountains a few miles east of the crest. The sections of Fig. 11.19 show the thick succession of folded beds from Cambrian to Devonian.

The axis of the Green Mountain anticlinorium trends north-northeast across the northwest corner of the Hyde Park quadrangle. This anticlinorium, which

is the principal structural feature of the bedrock of Vermont, extends north-northeast from the Massachusetts-Vermont border the full length of the state and about 50 miles into Quebec, a total distance of about 210 miles. The stratigraphic sequence and lithologic character of the rocks on the west limb of the anticlinorium are different from those on the east limb, and a generally accepted correlation of the two is not yet possible. In the Hyde Park quadrangle, and in the Montpelier quadrangle (Cady, 1956), which borders on the south edge of the Hyde Park quadrangle, the general eastward dip of the rocks is interrupted by a group of anticlines whose axes parallel the axis of the Green Mountain anticlinorium. [See Fig. 11.18.]

The bedrock of the quadrangle(s) comprises chiefly metamorphosed sedimentary and volcanic rocks, principally schist, phyllite, slate, granulite, quartzite, greenstone, amphibolite, crystalline limestone, and conglomerate, that range in age from Cambrian probably to Devonian. Intrusive igneous rocks, some of which are metamorphosed, underlie less than 1 percent of the area and comprise serpentinite and its derivatives (talc-carbonate rock and steatite), granite, and diabase that range in age from Ordovician probably to Mississippian.

All the rocks in this erea except the lamprophyre dikes have been affected by regional metamorphism. In this area, chlorite, garnet, and kyanite have been interpreted as successively general indicators of increasing metamorphic grade in the schists. Similarly, chlorite, actinolite, and hornblende are indicators in the greenstone and amphibolite. Most of the Hyde Park quadrangle is in the chlorite zone of metamorphism (Cady, 1956).

Bodies of serpentinite or its alteration products, talc carbonate rock and steatite, are numerous, having been noted in fifteen places by Albee and in five by Cady. They occur chiefly in the Stowe formation.

The serpentinite (or its derivatives) forms tabular, lenticular, or pod-shaped masses that strike north-northeast and dip steeply, parallel with the schistosity and commonly also with the bedding of the enclosing rocks. The serpentinite is dark green to dark greenish black on the fresh surface but weathers to a characteristic pale greenish-white or light-buff rind traversed by a reticulate system of sharply cut lines; it is composed almost entirely of the mineral serpentine, probably of the antigorite variety. The talc-carbonate rock is mottled greenish gray and weathers brown; it is composed of the minerals talc, magnesite, and locally small amounts of dolomite. The steatite ranges from white to green and greenish gray and weathers grayish tan; it is composed of the mineral talc (Albee, 1957).

Thick sills of granite invade the Waits River formation of the Montpelier quadrangle, and have generated cordierite and diopside as contact metamorphic effects. These sills are probably a late element of the Acadian folding which took place in Mid- and Late Devonian time (Cady, 1956).

The minor folds do not accord with the major folds.

The axes of most of the minor folds and granular quartz columns, as well as the intersections of fold bands and of slip-cleavage lamellae with bedding, are nearly vertical. This attitude implies that most of these minor structural features were not produced by shearing movements in a nearly east-west oriented vertical plane, such as were evidently responsible for the gently plunging structures of the Green Mountain anticlinorium. Instead they were probably either formed before folding of the anticlinorium by shearing movements in a north-south vertical plane, or after folding and tilting of the limbs of the anticlinorium by shearing movements in a north-south vertical plane, or after folding and tilting of the limbs of the anticlinorium to near vertical, by shearing movements in a horizontal plane. The pattern of movement of these minor folds is uniform over rather wide areas; thus most of the folds in the fold bands in the Moretown formation southeast of the Worcester Mountains are dextral in plan (see White and Jahns, 1950, p. 197, for usage of terms "dextral" and "sinistral"), and it appears that the rocks to the east have moved south relative to those to the west. This relationship is well shown at the previously cited exposures of the Moretown formation in Middlesex Gorge (Albee, 1957).

Central and East-Central Vermont

The outcrop pattern of three key formations in central and eastern Vermont is broadly shown on the map of Fig. 11.20, and the stratigraphic succession in Fig. 11.18. According to White and Jahns:

The formations of central and east-central Vermont are exposed as a series of parallel belts that strike nearly north. Most of the rocks dip steeply, and many are overturned. With one possible exception, there seem to be no major repetitions within the sequence, and the order of formations from west to east appears to be the same as the order of their deposition. The formations are dominantly schist or phyllite, with varying proportions of arenaceous material. One thin formation, the Shaw Mountain, contains quartz conglomerate, calcareous tuff, and crinoidal limestone. The third-from-highest formation, the Waits River, is very thick and contains a large proportion of calcareous beds. The distance from the base of the lowest formation to the top of the highest, measured normal to bedding, is more than 100,000 feet; this large apparent thickness is believed to be not very much greater than the original thickness.

The metasediments have been intruded by granitic dikes and plutons, mafic dikes, and small ultramafic plutons.

Two principal stages of deformation are distinguished. During the earlier stage the rocks were folded, and a schistosity was developed nearly parallel to bedding. Throughout the area the minor folds of this stage indicate a consistent upward movement of rocks on the east with respect to those on the west. The folds plunge at low to moderately steep angles, typically northward.

Phenomena associated with the later stage of deformation decrease in intensity both eastward and westward from the belt underlain by the calcareous Waits River formation. At a distance from this formation, the rocks have prominent slip cleavage, and the earlier schistosity is folded. The minor folds plunge moderately to steeply northward on the western side of the area and more gently northward on the eastern. As the Waits River formation is approached, slip cleavage passes gradually into a schistosity that obliterates the earlier schistosity, and the intensity of later folding increases. In both the eastern and the western parts of the area the later minor folds indicate that the rocks of the Waits River formation have moved upward with respect to the formations on either side.

The central part of the belt underlain by the Waits River formation is marked by a huge arch, 10–20 miles across, whose axis is more or less parallel to the belt and plunges gently northward. This is shown to be an arch, not in bedding, but in the later schistosity and in the axial planes of large isoclinal folds that were formed during the later stage of deformation. The axial planes of three of these large isoclinal folds can be correlated across the crest of the cleavage arch at Strafford Village.

Western, Central, and Northern New Hampshire

Stratigraphy. A series of metasedimentary and metavolcanic rocks in western, central, and northern New Hampshire ranges in age from Ordovician (?) to Lower Devonian and has an aggregate thickness of 16,000 feet. See Fig. 11.18. Figure 11.21 is a columnar section of the Littleton–Moosilauke area in the White Mountains of west central New Hampshire. The stratified rocks fall into six major units. The Albee, Ammonoosuc, and Partridge formations are of pre-Silurian, probably Upper Ordovician age, the unconformably overlying beds are the Clough conglomerate and Fitch formation of Silurian age, and the Littleton formation is of Lower Devonian age. The Albee was originally a shale and sandstone formation, and although no fossils have been found in it, it appears to be above the fossiliferous Middle Ordovician of Vermont (Billings, 1937).

The Ammonoosuc volcanics consist principally of soda-rhyolite, soda-rhyolite volcanic conglomerate, meta-andesite porphyry breccia, and slate and impure quartzite. The Partridge formation is largely a black slate. In

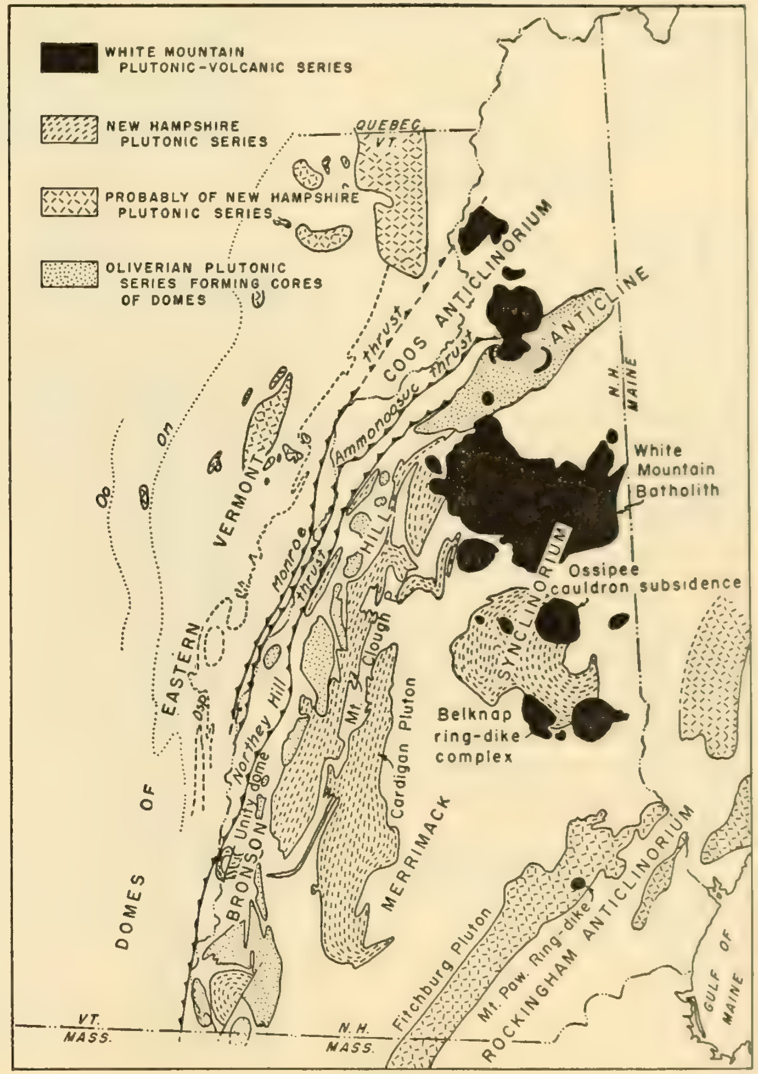


Fig. 11.20. Major structures of eastern Vermont and New Hampshire. After White and Johns (1950) and Billings (1956). The narrow Connecticut Valley synclinorium lies between the Northey Hill and Ammonoosuc thrusts and is not labeled on the map. Osp, Standing Pond volcanics; On, Northfield slate; Oo, Ottauquechee phyllite.

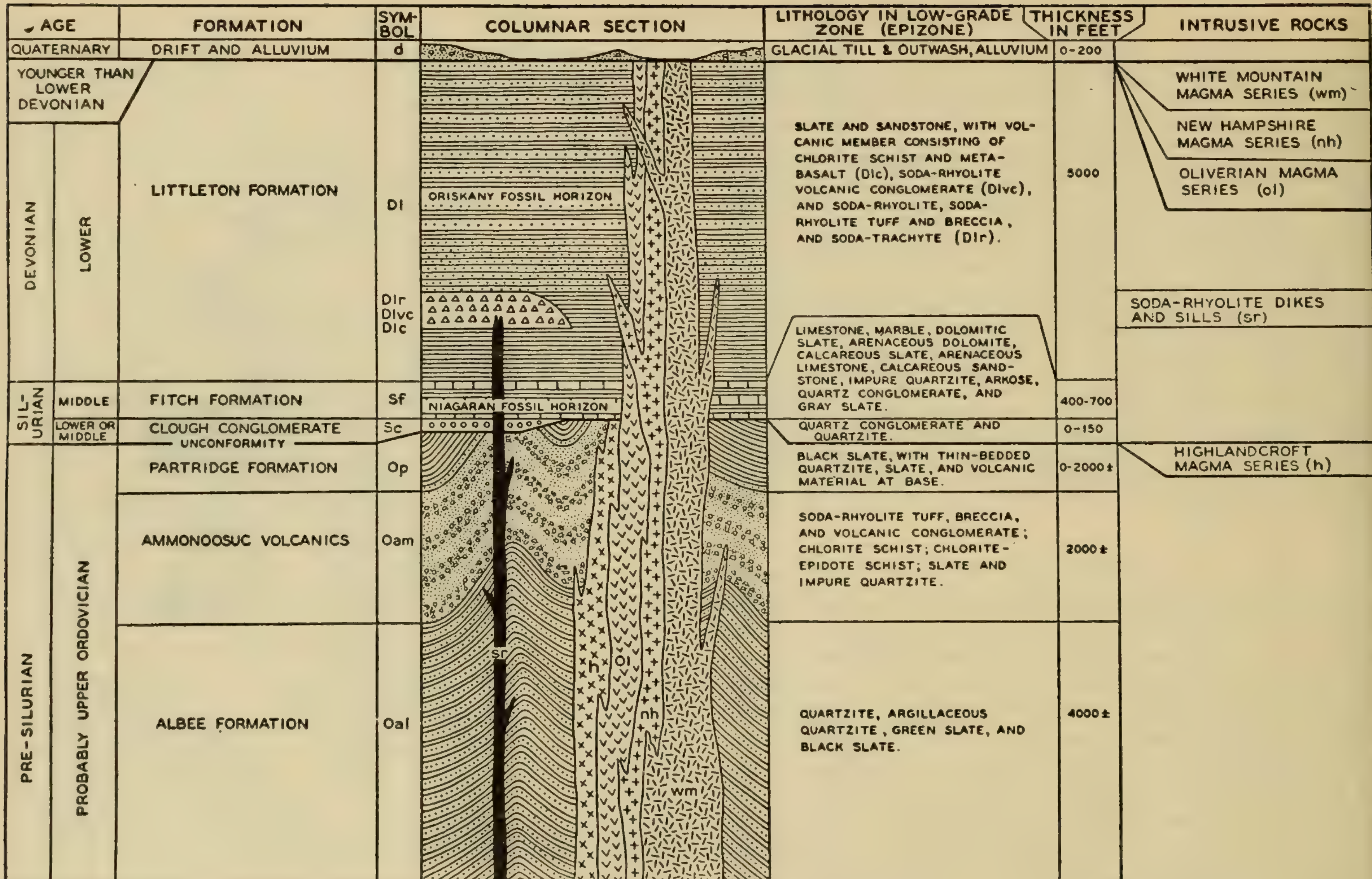


Fig. 11.21. Columnar section of the Littleton-Moosilauke area. Reproduced from Billings, 1937. In addition to the sequence and character of the sedimentary and volcanic rocks, the time of intrusion of igneous rocks is shown.

places at the base, black slate and fine-grained, light quartzite alternate in beds a quarter of an inch thick.

The Clough conglomerate is one of the best key horizons in western New Hampshire, and although thin, it is resistant and exceptionally well represented in outcrops. It apparently continues southward to the Massachusetts boundary. Its outcrops are generally white cliffs. The pebbles in the conglomerate are chiefly vein quartz, but some are quartzite, jasper, greenstone, or soda-rhyolite. In places only a few pebbles are present; in others they constitute over 60 percent of the rock (Billings, 1937). The matrix is pure or slightly impure quartzite.

> The Clough conglomerate directly underlies the Fitch formation which carries middle Silurian fossils. Moreover, the two formations are closely related in age, for a few beds of quartz conglomerate are found in the Fitch. The Clough conglomerate, however, is separated from the underlying strata by an unconformity. It is apparent that the formation is either middle or lower Silurian. In many respects the Clough is similar to the Shawangunk conglomerate of New York, although the former is thinner and purer. The Clough underlies fossiliferous middle Silurian, and the Shawangunk carries middle Silurian fossils in its upper part. The two are closely related, if not identical, in age (Billings, 1937).

The Fitch formation in its least altered form consists of white to buff marble; gray limestone and marble; buff dolomitic slate; buff to brown arenaceous dolomitic limestone; gray calcareous slate ("trilobite slate" of earlier workers); white to gray arenaceous limestone and calcareous, arkosic conglomerate; gray impure quartzite; white to gray arkose; white quartz conglomerate; and gray slate. Fossils have been found at two localities in the Fitch formation southeast of the Ammonoosuc thrust, and are recognized as of Middle Silurian (Niagaran age).

The Littleton formation of Lower Devonian age consists in its least metamorphosed condition chiefly of slate and sandstone, with subordinate amounts of soda-rhyolite conglomerate, tuff and breccia, and some greenstone.

Formations older than those listed in the chart of Fig. 11.21 are known. The Orfordville formation, first recognized in west central New Hampshire (Kruger, 1946) underlies the Albee formation, and the Waits River formation first found in central Vermont, underlies the Orfordville (Currier and Jahns, 1941). The base of the Waits River is 2000 feet above a crinoidal limestone which appears to be Middle Ordovician. If so, both the Waits River and Orfordville are Middle Ordovician or younger. The Orfordville formation was originally a shale with very thin beds of sandstone, and the Waits River a calcareous shale and limestone formation.

Structure

General Statement. In Massachusetts and southern New Hampshire the structures trend northerly; in northern New Hampshire they veer northeasterly. A succession of anticlinoria and synclinoria make up the major elements of the structure. See Figs. 11.20 and 11.26. Proceeding eastward from the great monocline of central and eastern Vermont three thrust faults occur, and between the middle (Ammonoosuc) and eastern (Northey Hill) is the Connecticut Valley synclinorium. This lies approximately astride the boundary line of Vermont and New Hampshire. Next east is the Bronson Hill anticline, the Merrimack synclinorium and in southeastern New Hampshire the Rockingham anticlinorium. The Coos anticlinorium is in the northern part of the state and lies between the Monroe and Ammonoosuc thrusts.

The older plutonic series, especially the Oliverian and New Hampshire series, participate in the northerly and northeasternly trend. This may be seen by the Oliverian series making up the cores of the domes along the Bronson anticline, and by the foliated Mt. Clough and Cardigan plutons of the New Hampshire series striking along the western flank of the Merrimack synclinorium.

Bronson Hill Anticline. The Bronson Hill anticline extends from Massachusetts to Maine, a length of 150 miles. It ranges from 6 to 16 miles wide. The core is composed of the Ammonoosuc volcanics and the Oliverian plutons with the Clough, Fitch, and Littleton formations on both flanks.

Rockingham Anticlinorium. The Rockingham anticlinorium, lies in southeastern New Hampshire, between the Atlantic Ocean and the Fitchburg pluton. The individual folds of the anticlinorium are, from southeast to northwest, the Rye anticline, the Great Bay (Eliot) syncline, and the Exeter anticline (largely occupied by the Exeter pluton).

Merrimack Synclinorium. East of the Bronson Hill anticline and northwest of the Rockingham anticlinorium is a large area of Littleton formation, all in the sillimanite zone of metamorphism. Inasmuch as this band of the Littleton formation is bordered on either side by older strata, it must occupy a synclinorium. This structural feature is called the Merrimack synclinorium, because much of it is drained by the Merrimack River and its tributaries.

Throughout much of western New Hampshire the western limb of the Merrimack synclinorium is invaded by large bodies of the New Hampshire plutonic series. These relations are well shown on sections A–A′ and B–B′–B″ of Fig. 11.26.

Thrust Faults. The Ammonoosuc thrust is marked generally by Ammonoosuc volcanics being thrust over the Littleton formation with a stratigraphic displacement of 7000 feet. The fault dips from 32 to 50 degrees westerly. It is younger than the regional metamorphism.

The Northey Hill thrust predates the metamorphism because there is no break in grade of metamorphism across it. This feature renders recognition of the fault a little difficult, yet mapping shows the Littleton formation lies in contact with several different formations along it, and a maximum stratigraphic displacement of 12,000 feet may be measured. A steep dip characterizes much of its length, and this is believed due to later deformation.

The Monroe thrust is about as long as the Ammonoosuc (85 miles). It is nearly vertical throughout most of its length, but in places dips southeasterly. It is mostly older than the regional metamorphism, but later deformation steepened it and also caused some renewed movements along it.

Magma Series

Plutonic rocks are abundant and varied in form and composition. Four magma series have been worked out (Billings, 1937). The oldest is known as the Highlandcroft magma series and is probably of late Ordovician age. See chart, Fig. 11.21 and map, Fig. 11.20. Some time after the Lower Devonian, probably in Mid- and Late Devonian time, other large quantities of magma invaded the region. The Oliverian magma series preceded the folding and was followed by the New Hampshire magma series, the earlier members of which were contemporaneous with the main period of folding, and the later members of which were slightly younger than the folding. The White Mountain magma series is the youngest of the plutonic rocks, and it appears less extensive than the others. It is probably early Mississippian in age (Billings, 1945).

The Highlandcroft magma series is represented by the Highlandcroft granodiorite and small bodies of diorite, quartz diorite, and quartz monzonite. The Oliverian magma series is represented by the pink Owls Head granite in the Littleton area and by other units in the Rumney, Mt. Cube, and Mascoma quadrangles. Many sills in the Ammonoosuc volcanics are of this series.

The White Mountain magma series is characterized by ring-dikes, stocks, a batholith, and by eruptive differentiates. According to Billings, 1945:

> Much of the magma of the White Mountain magma series was erupted on the surfaces to from the Moat volcanics. Tuffs, breccias, and lavas, composed chiefly of rhyolite, andesite, and basalt, but also including some trachyte, are typical. Rhyolite is by far the most common; trachyte is rare.
>
> The intrusive rocks range in composition from gabbro to granite, and a great variety of intermediate types are developed. Chapman and Williams, in a careful, detailed study, have shown that the mafic rocks are the oldest and the felsic are the youngest. They have also determined the areal extent of the plutonic rocks and calculated the percentage of each compared to the whole magma series. The order of intrusion, from oldest to youngest, and the percentage of each as exposed at the surface, are gabbro, norite, diorite, and quartz diorite (0.5 per cent); monzodiorite and monzonite (1.5 per cent); syenite, including some nepheline-sodalite syenite (9 per cent); quartz syenite (10 per cent); granite and granite porphyry (79 per cent). Although the rocks in general became more siliceous as differentiation progressed, this is not true in detail. Especially important is the fact that the Albany quartz syenite is younger than the granite porphyry. This is significant in considering the tectonic evolution of the area.
>
> Chapman and Williams have also shown that fractional crystallization controlled the evolution of the series, but that abyssal assimilation played an important role.
>
> The Moat volcanics, in large part contemporaneous with the granite porphyry, are older than the Albany type of quartz syenite, but their age relative to the more mafic plutonic rocks is uncertain.

Metamorphism

All the sedimentary and metamorphic rocks have been deformed and metamorphosed to various degrees. The metamorphism increases generally to the southeast, and three zones have been recognized by Billings, namely, the low-grade, the middle-grade, and the high-grade. See map of Fig. 11.22.

The distinction between the zones is based primarily on their mineralogy. The low-grade zone is characterized by chlorite, epidote, albite, sericite, and dolomite; the middle-grade zone, by staurolite, garnet, hornblende, actinolite, diopside, biotite, and intermediate and calcic plagioclase. The mineralogical contrast between these two zones is striking. The high-grade zone differs from the middle-grade zone chiefly in that sillimanite is present and staurolite is absent or is in small crystals. Thus, if aluminous sediments are not present, it is difficult or impossible to distinguish the middle-grade and the high-grade zones on mineralogical criteria alone. In general, the high-grade rocks are coarser than the middle-grade, but this criterion is difficult to apply, and, wherever the rocks might belong to either of the two higher zones, they have been assigned to the middle-grade zone.

The change in the degree of metamorphism in a southeasterly direction is readily apparent. The cumulative effect of these changes is so great that, for a long time, rocks now known to belong to the same formations were believed to be of very different ages. Whereas, northwest of the Ammonoosuc thrust the rocks are dominantly sandstone, slate, calcareous slate, dolomitic slate, rhyolite tuff, and greenstone, composed of such minerals as sericite, chlorite, albite, dolomite, calcite, quartz, and epidote, to the southeast the rocks are mica schist, calcite-biotite schist, actinolite-diopside granulite, biotite gneiss, and amphibolite, composed of such minerals as biotite, garnet (almandite), staurolite, sillimanite, actinolite, diopside, hornblende, calcite, quartz, and calcic plagioclase. Moreover, there is a general coarsening in grain. These changes clearly represent progressive metamorphism toward the southeast, for the new rocks are farther and farther removed mineralogically from the original rocks from which they were derived.

A number of the intrusive rocks are older than the regional metamorphism and were affected to different degrees. The Highlandcroft granodiorite was in the zone of low-grade metamorphism, and its original andesine plagioclase has been replaced by albite-oligoclase, epidote, and sericite. Green biotite, which is found in places as a shell around the hornblende, is of metamorphic origin. The Moulton diorite has been subjected to low-grade metamorphism, and its original condition is much altered.

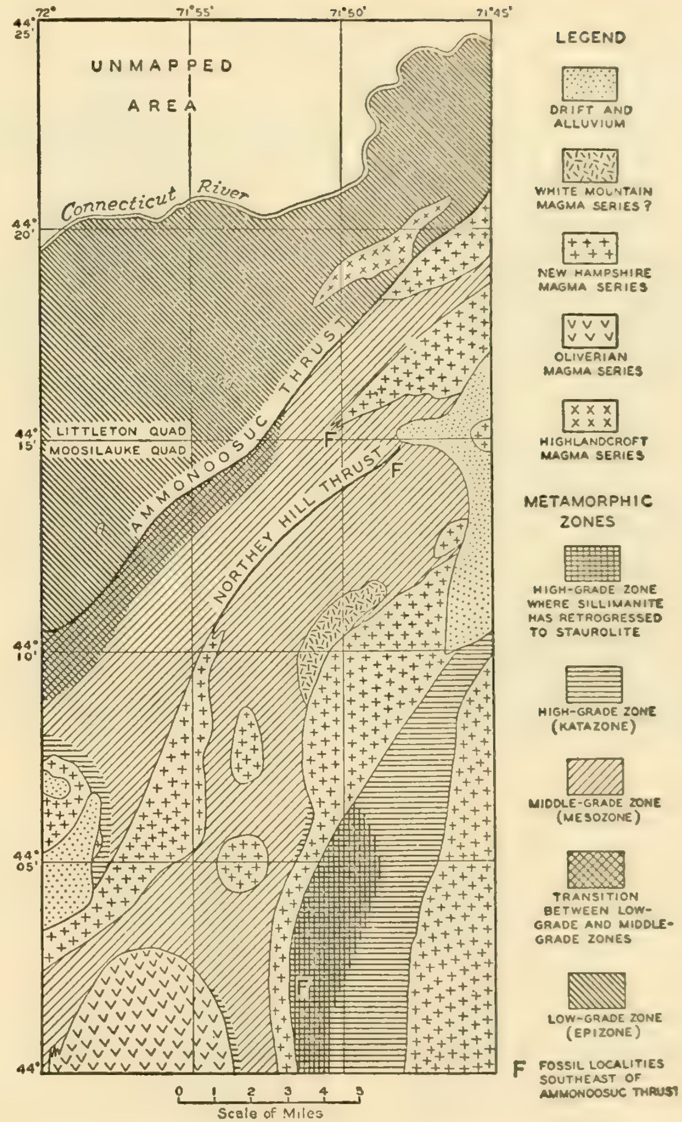


Fig. 11.22. Metamorphic zones in the Littleton-Moosilauke area. Metamorphism is progressive toward the southeast. Reproduced from Billings, 1937.

Basic dikes and sills have attained equilibrium under the new metamorphic conditions.

Billings regards the main alteration to have occurred after the Northey Hill thrust and during the intrusions of the New Hampshire magma series. Then the Ammonoosuc thrust brought different metamorphic zones into sharp contact with each other. Also in certain places retrograde metamorphism set in with the formation of much chlorite.

The cause of the metamorphism is apparently the intrusions of the various plutons of the New Hampshire magma series. Northwest of the Ammonoosuc thrust where metamorphism is least, the intrusions of the New Hampshire magma series are absent except that a few small bodies of the Bethlehem gneiss and Kinsman quartz monzonite appear. Billings points out that, as intrusions are common eastward to the Maine border, and as the sedimentary rocks almost invariably are recrystallized to high-grade metamorphic rocks, there must be a causal connection between the increase in metamorphism and these intrusions. Not only is there a general increase in the intensity of metamorphism toward the area where igneous intrusions are most abundant, but there is an increase locally toward individual bodies. Such high-grade zones surrounding intrusive masses are not well defined in the map of Fig. 11.22, but it is suggested that the contact metamorphic zones vary in width greatly, and that certain zones betray the presence of unexposed plutons.

Mechanics of Instrusion

Introduction. The post-tectonic White Mountain magma series is characterized by ring-dikes, stocks, and a batholith (Billings, 1945). The ring-dikes, most of which range in composition from monzonite to quartz syenite, intruded arcurate and circular vertical fracture zones by piecemeal stoping and related mechanisms. Cauldron subsidence, although associated with some ring-dikes, is not essential for their intrusion. The stocks of the White Mountain magma series were emplaced by underground cauldron subsidence.

The New Hampshire magma series, emplaced during the Acadian orogeny, occurs chiefly as great sheets, lenses, and stocks, forcefully injected into the older formations.

Ring-Dikes. Altogether, 36 ring-dikes associated with the White Mountain magma series have been discovered in New Hampshire. A ring-dike complex is a structural unit containing one or more ring-dikes. According to Billings (1945):

> There are five ring-dikes at Mt. Tripyramid, four each in the Pliny region and the Franconia quadrangle, and six in the Belknap Mountains, although the six separate intrusions could be considered to belong to two composite ring-dikes. Ring-dikes have also been described from adjacent areas in Quebec and Maine.
>
> Complete ring-dikes that encompass 360 degrees are rare, but the ring-dike of the Ossipee Mountains and some of those on Mt. Tripyramid are of this type. Most ring-dikes are arcuate in plan and those in New Hampshire encompass, on the average, 170 degrees of the total possible 360 degrees. The average radius of ring-dikes in New Hampshire, measured from the outer margin of the ring-dike to its center of curvature, is three miles. A ring-dike composed of Albany quartz syenite in the Franconia quadrangle has a radius of 9.2 miles and is one of the largest known anywhere in the world. The smallest ring-dike in New Hampshire, with a radius of only 0.8 mile, is on Mt. Tripyramid. The average width of ring-dikes in New Hampshire is 1900 feet. The arcuate body of amphibole granite in the southern part of the Franconia quadrangle is 14,000 feet wide, but this may not be a true ring-dike.

Inside some of the ring-dikes are accumulations of extrusive rocks, known as the Moat volcanics. They are never found outside the ring-dike. The volcanics also have the same composition as the ring-dike within which they have subsided.

> The Moat volcanics are at least 10,000 feet thick and rest with pronounced angular unconformity on the older metamorphic rocks of the Littleton formation and the plutonic rocks of the New Hampshire magma series. It is almost always impossible to determine the attitude of the Moat volcanics, because many of the pyroclastic rocks and lavas are devoid of bedding and flow structure. Available data indicate, however, that near the ring-dikes the volcanics are essentially vertical, but toward the center of the complex the dips become progressively less [Fig. 11.23].
>
> Unfortunately, precise data concerning the amount of subsidence are difficult to obtain in New Hampshire. The key horizon used for such studies is the base of the Moat volcanics. It is apparent from Fig. 11.23 that the center of the subsided block has settled 10,000 feet relative to the margins of the block near the ring-dike. Moreover, the edge of the subsiding block just inside the ring-dike has apparently settled at least 5,000 feet relative to the rocks some distance outside of the ring-dike. Therefore, the center of the subsided block has dropped at least 15,000 feet relative to the rocks some distance outside of the ring-dike.

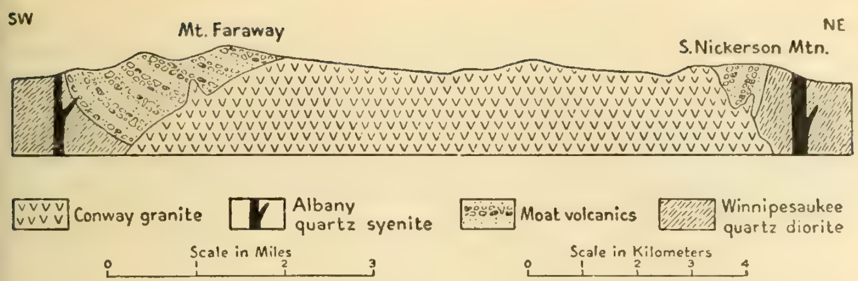


Fig. 11.23. Section through the Ossipee Mountains, N. H. Reproduced from Billings, 1945, after Kingsley.

It is apparent that the intrusion of some ring-dikes is associated with the subsidence of a central block. It does not follow, however, that all ring-dikes are associated with central subsidence.

Billings (1945) believes, because the ring-dikes are vertical in New Hampshire, that their intrusion was controlled by an annular vertical fracture zone, the width of which was comparable to the width of the ring-dike. Such a fracture zone would be susceptible to piecemeal stoping. Various combinations of the annular or partially annular fracture zone with sagging or doming are shown in Fig. 11.24.

Stocks. For most of the stocks there are few data to indicate whether they are concordant or discordant because many of them have been intruded into areas already occupied by relatively massive or weakly foliated older plutonic rocks. The Mt. Ascutney stock has been shown to cut discordantly across the steeply dipping older strata, and the lineation and fold axes of the older strata have not been modified by the intrusion (Chapman and Chapman, 1940). A process of underground cauldron subsidence, whereby large blocks with outward-dipping walls approximately the size of the present stocks sank, is visualized, and is illustrated in Fig. 11.25. The activity occurred in the last stages of the evolution of the White Mountain magma series. The remarkable uniformity of the White Mountain magma series through New Hampshire suggests that a single reservoir underlay much of the state (Billings, 1945).

Plutons of Forceful Injection. Many plutons belonging especially to the New Hampshire magma series have been emplaced by forceful in-

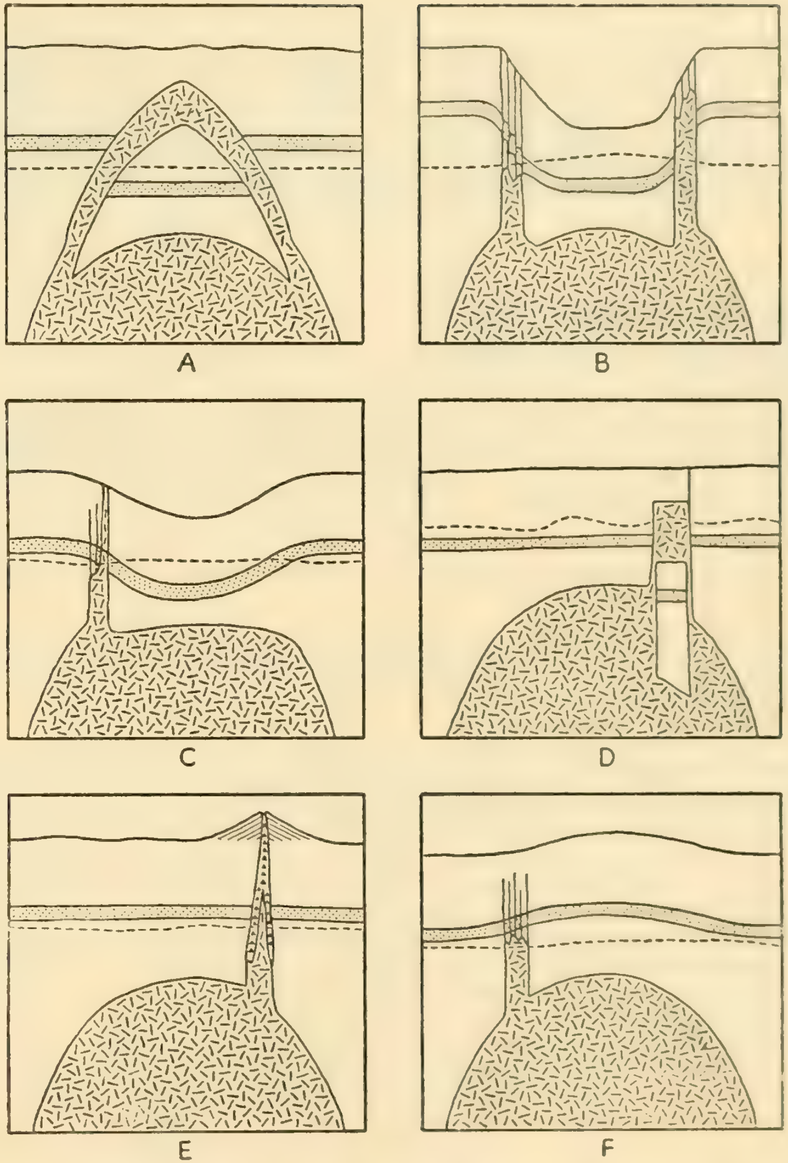


Fig. 11.24. Origin of ring-dikes. Reproduced from Billings, 1945. Broken line is present erosion surface.

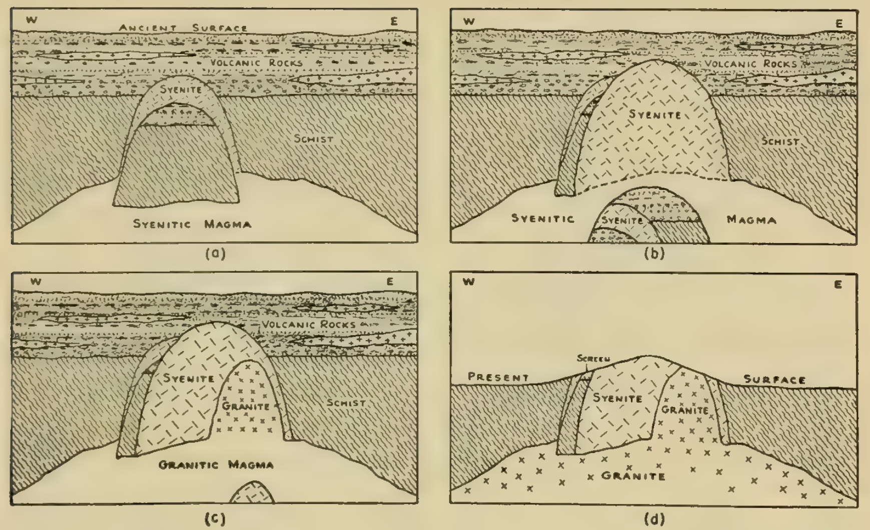


Fig. 11.25. Evolution of the syenite-granite stock of Ascutney Mountain, Vt. Reproduced from Chapman and Chapman, 1940.

jection. Notable of these are the Kinsman quartz monzonite and the Bethlehem gneiss.

According to Billings (1945):

The Mt. Clough pluton, composed of Bethlehem gneiss, is undoubtedly the longest intrusion in New Hampshire. The main body extends southward for 90 miles from the northern part of the Franconia quadrangle to the south end of the Lovewell Mountain quadrangle, which is beyond the limits of Fig. 11.20. The width ranges from half a mile to 7 miles. In the Moosilauke quadrangle the contacts are essentially vertical and the pluton is a vertical sheet. Further south, however, the contacts dip to the east and along the eastern border of the Mascoma quadrangle and the western border of the Cardigan quadrangle, the upper and lower contacts dip 30 degrees east. Here the pluton is a huge sheet inclined to the east [Fig. 11.26].

A series of plutons composed of Kinsman quartz monzonite lie east of the Mt. Clough pluton. The most northerly of these, which may be called the Kinsman pluton . . . is a gigantic lens, essentially vertical in the surrounding schists.

In the western part of New Hampshire, some ten miles east of the Connecticut River, the crest of a major anticline is occupied by a series of "domes." In their essential features these domes, nine of which have been mapped, are remarkably similar. A central oval-shaped core of plutonic rocks, ranging in composition from granodiorite through quartz monzonite to granite, has a foliation that dips outward. The plutonic rocks, overlain by Ordovician (?), Silurian, and Devonian strata, include the Ordovician (?) rocks in many localities and the Silurian rocks in at least one locality. The upper contact of the plutonic rocks is at essentially the same stratigraphic horizon in all the domes, approximately 500 feet below the top of the Ammonoosuc volcanics, but ranges from the top to an horizon 1,000 feet below the top. The overlying formations likewise participate in the domical structure.

Originally considered to be laccoliths or "bottomless" plugs that had bowed up their roof, it is possible that they all belong to a single great concordant sheet, originally horizontal, that has been buckled up during orogeny.

Tectonic History

Ordovician Sedimentation. The oldest rocks known so far in the eugeosyncline of New Hampshire are Middle Ordovician limestone, calcareous shale, and shale, 7000 to 8000 feet thick. Over these accumulated the Upper Ordovician Ammonoosuc volcanics, about 4000 feet thick, and over the volcanics another 500 to 2000 feet of shale.

Taconic Orogeny. Near the close of Ordovician time the previously deposited sediments and volcanics were mildly folded and eroded. The disturbance here probably marks the subdued effects of the Taconic orogeny of the Hudson-Champlain region farther west.

Silurian and Lower Devonian Sedimentation. In a Middle Silurian sea that moved in from the southwest, conglomerates and sands of the Clough formation and the dolomitic sandstones and shales of the Fitch formation, not over 800 feet thick, were deposited. Late Silurian history is obscure, but during early Devonian time about 10,000 feet of sandstone, shale, and volcanic materials accumulated. See upper left section of Fig. 11.27.

Acadian Orogeny. During Mid- or Late Devonian, the strata were caught in a major orogeny. Even before the deformation, or at least in its early stages, successive injections of the Oliverian magma series formed a great sheet in the Ammonoosuc volcanics, later to be domed in several places along the western margin of New Hampshire. The Ordovician, Silurian, and Devonian strata were thrown into a series of anticlinoria and synclinoria whose axes trend north and northeast, and countless minor folds were impressed upon the larger. Also the Northey

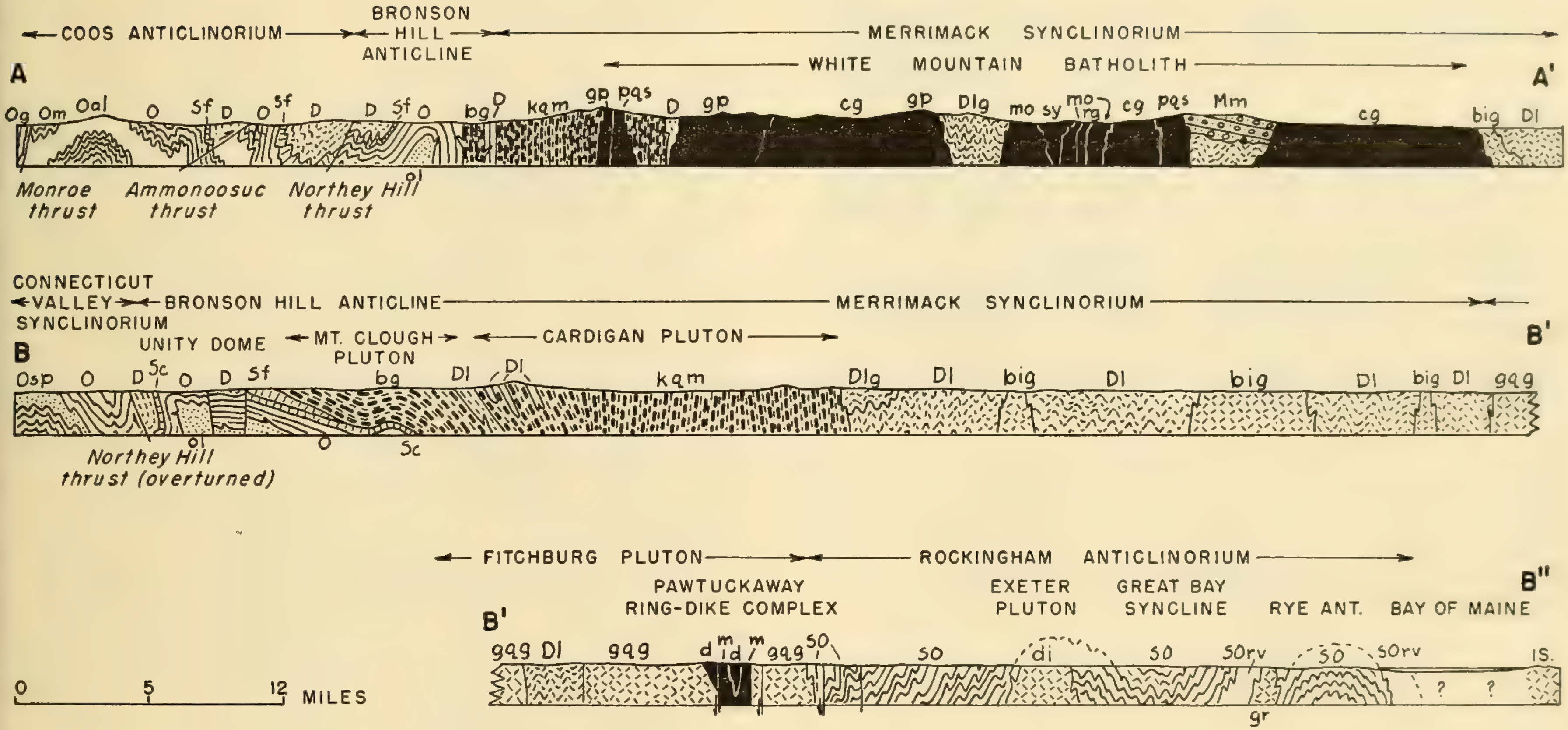


Fig. 11.26. Cross section of New Hampshire. After Billings, 1956. Section A–A′ is across northern part of state and B–B′–B″ across southern part. Refer to map, Fig. 11.20.

Hill thrust occurred. Schistosity parallel to the bedding formed during the earlier stages of the folding, and fracture cleavage, essentially parallel to the axial planes of the minor folds, formed during the later stages. The rocks were subjected to low-grade metamorphism northwest of the Ammonoosuc thrust, and to medium and high-grade alteration southeast of it. The main metamorphism occurred after the Northey Hill thrust and during the intrusions of the New Hampshire magma series which were chiefly responsible for the medium- and high-grade metamorphism. See third section in Fig. 11.27.

Succeeding the metamorphism was the Ammonoosuc thrusting and, following this, some normal faulting. Then the Moat volcanics were erupted, and the plutons of the White Mountain magma series were emplaced to complete the bedrock complex. This may have occurred in Mississippian time. Examine the last four diagrams of Fig. 11.27.

Isotope Ages and the Acadian Orogeny

It is becoming evident that the Devonian period began almost 400 m.y. ago, and that our previous estimates that designate this age for

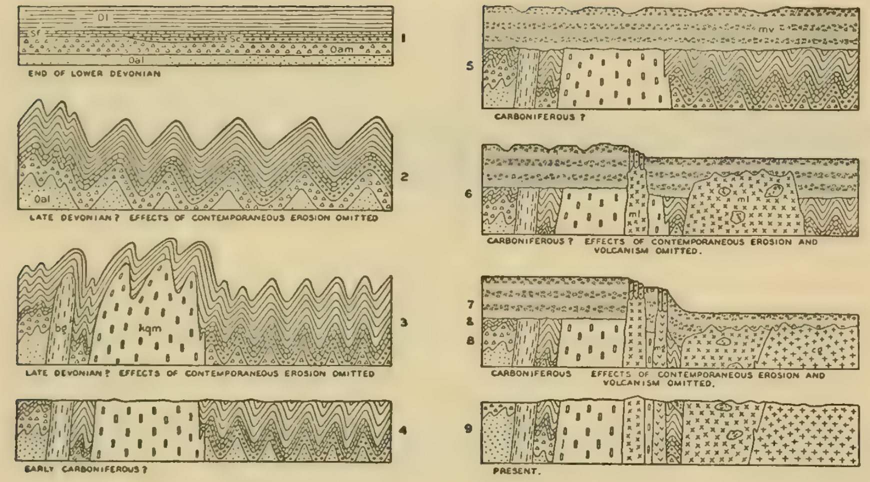


Fig. 11.27. Evolution of the Franconia quadrangle terrane, White Mountains, N. H. Reproduced from Williams and Billings, 1938. Oal, Albee formation; Oam, Ammonoosuc volcanics; Sc, Clough conglomerate; Sf, Fitch formation; Dl, Littleton formation; bg, Bethlehem gneiss; kqm, Kinsman quartz monzonite; mv, Moat volcanics; ml, Mt. Lafayette granite porphyry; mq, Mt. Garfield prophyritic quartz syenite; eg, Conway granite and Mt. Osceola granite. Bethlehem and Kinsman belong to the New Hampshire magma series.

the Late or Mid-Ordovician must be revised. Hurley *et al.* (1959) report the age of a quartz monzonite stock in northwestern Maine which intrudes well-documented, fossiliferous, Lower Devonian slate as 360 m.y. The metamorphism of the beds is believed to have occurred along with the intrusion.

> Therefore, the Oriskany sedimentation took place prior to this time. This is in agreement with findings of Fairbairn in Nova Scotia where sediments of similar age have been intruded by granitic rocks . . . (Hurley *et al.* 1959).

Ages in the 320–380 m.y. range category have generally been correlated with the Taconic orogeny, but if they indicate Acadian orogeny, then we must conclude that nearly all the metamorphism and most of the plutonic activity is Acadian in New England and the crystalline Piedmont.

CARBONIFEROUS BASINS

Location

Emerson in 1917 recognized five major Carboniferous basins and a number of minor ones in eastern Massachusetts, southeastern New Hampshire, and Rhode Island, and they are shown on the Geological Map of the U.S. (1932) accordingly. The new geological map of New Hampshire by Billings (1956), however, recognizes the "Carboniferous" basins of Emerson in New Hampshire as Devonian and older, and therefore it appears that only two major basins are now to be considered, the Narragansett and the Boston. Two smaller basins in northern Rhode Island also are definitely demonstrated, and they will be referred to as the Woonsocket basins, following Emerson. The above basins are shown on the map of Fig. 11.28.

The Carboniferous stratified rocks are in the slope from the New England upland to the Seaboard lowland and in the lowland itself.

Narragansett Basin

The generalized stratigraphy of the three basins shown on the map of Fig. 11.28 is illustrated on the correlation chart of Fig. 11.29. The igneous intrusive rocks are also shown. It will be noted that the basement complex consists of metamorphosed Precambrian sediments and intrusives and various Acadian intrusives. Some fossiliferous Lower Cambrian beds are known in eastern Massachusetts (Chute, 1950).

According to Emerson (1917) the strata of the Narragansett basin are in large part coarse clastics with an aggregate thickness of 12,000 feet. At the base is the Pondville quartz conglomerate, which is a coarse, white, granitic waste or arkose 100 feet thick. Above the Pondville is the Wamsutta group of dominantly red conglomerates, sandstones, shales, slates, and felsite flows, breccias, and conglomerates, some 1000 feet thick. Above these strata are the thick Rhode Island coal measures that include dominantly dark gray conglomerate, pebbly sandstone, sandstone and graywacke, shale, and coal beds. They contain the Odontopteris flora and insect beds, and are about 10,000 feet thick. Above the coal measures is the Dighton conglomerate of the northern field and the

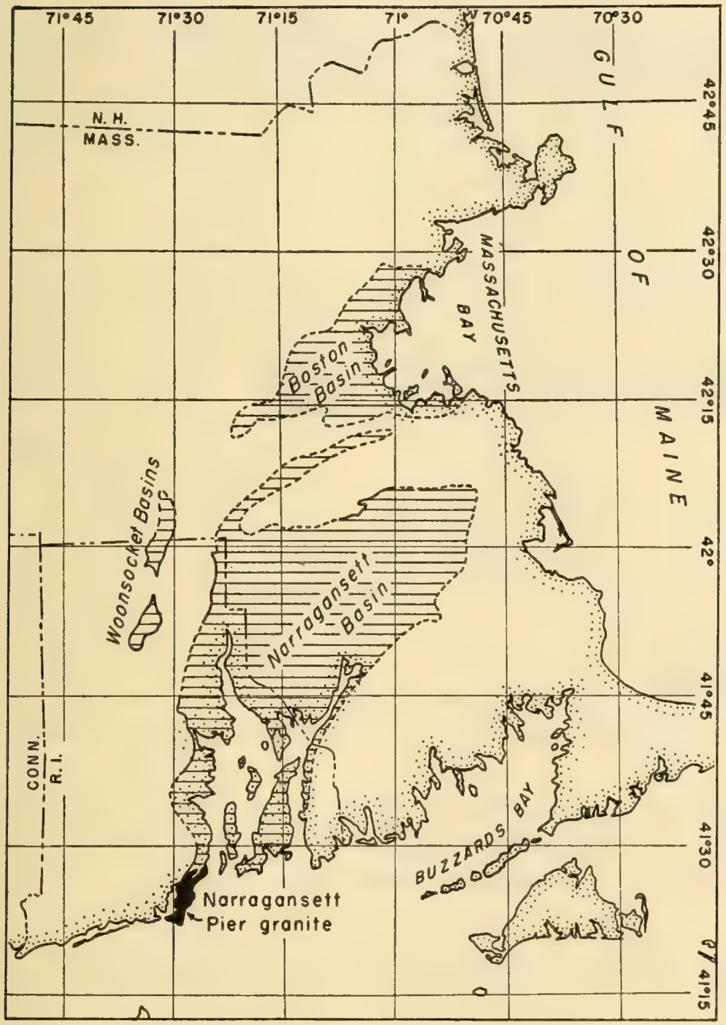


Fig. 11.28. Carboniferous basins of Rhode Island and Massachusetts.

Purgatory conglomerate of the southern field. The basin beds become metamorphosed to slates and quartzites to the south and the pebbles of the conglomerates are elongated and indented. They are regarded as Carboniferous in age and probably Pennsylvanian.

Recent detailed work by Richmond (1952) Quinn *et al.* (1949), Quinn (1951, 1952), Nichols (1956), Quinn and Springer (1954), and Chute (1950) is responsible for the correlation chart (Fig. 11.29), and the following generalizations. The succession of formations given by Emerson is not found in any one quadrangle. The unconformity at the base of the Pennsylvanian beds in the Narragansett basin is striking, and is shown by the near right angle discordance of the contacts of older formations with the Pennsylvanian, and by the discordance in outcrop of bedding and schistosity. Three episodes of metamorphism may be detected (Quinn, 1952). The beds of the Blackstone series were first moderately affected—sandstone to quartzite, mudstones to amphibolite schist. The later Esmond granite is mildly metamorphosed as are the volcanics of the East Greenwich group. Since the East Greenwich beds contain peb-

	NARRAGANSETT BASIN		WOONSOCKET BASINS		BOSTON BASIN	
	Sedimentary	Igneous	Sedimentary	Igneous	Sedimentary	Igneous
TRIASSIC		Diabase dikes		Diorite and diabase dikes		
PERMIAN?		Narragansett Pier granite & peg.		Pegmatite and aplite		
PENNSYLVANIAN	Dighton congl. Rhode Island fm. Wamsutta gr. Pondville congl.		Bellingham congl.		Cambridge slate Roxbury cgl. {Squantum tillite, Dorchester slate, Brookline cgl. & vols.}	
MISSISSIPPIAN		East Greenwich group (granite & vols.)				Quincy granite
DEVONIAN OR EARLIER		Esmond granite ?Scituate granite		Fine grained gran. Esmond granite Metadiabase dikes Scituate gr. gn.	Fossiliferous Lower Cambrian	Dedham granodiorite Salem gabbro-diorite Volcanic rocks
PRECAMBRIAN	Amphibolite schist Blackstone ser. {Sneech Pond schist, Westboro qtz., Mussey Brook schist}		Blackstone ser. Woonasquatucket fm. Absalona fm. Nipsachuck gneiss	Porphyritic metadiorite		

Fig. 11.29. Some sedimentary and igneous rocks of Rhode Island and Massachusetts.

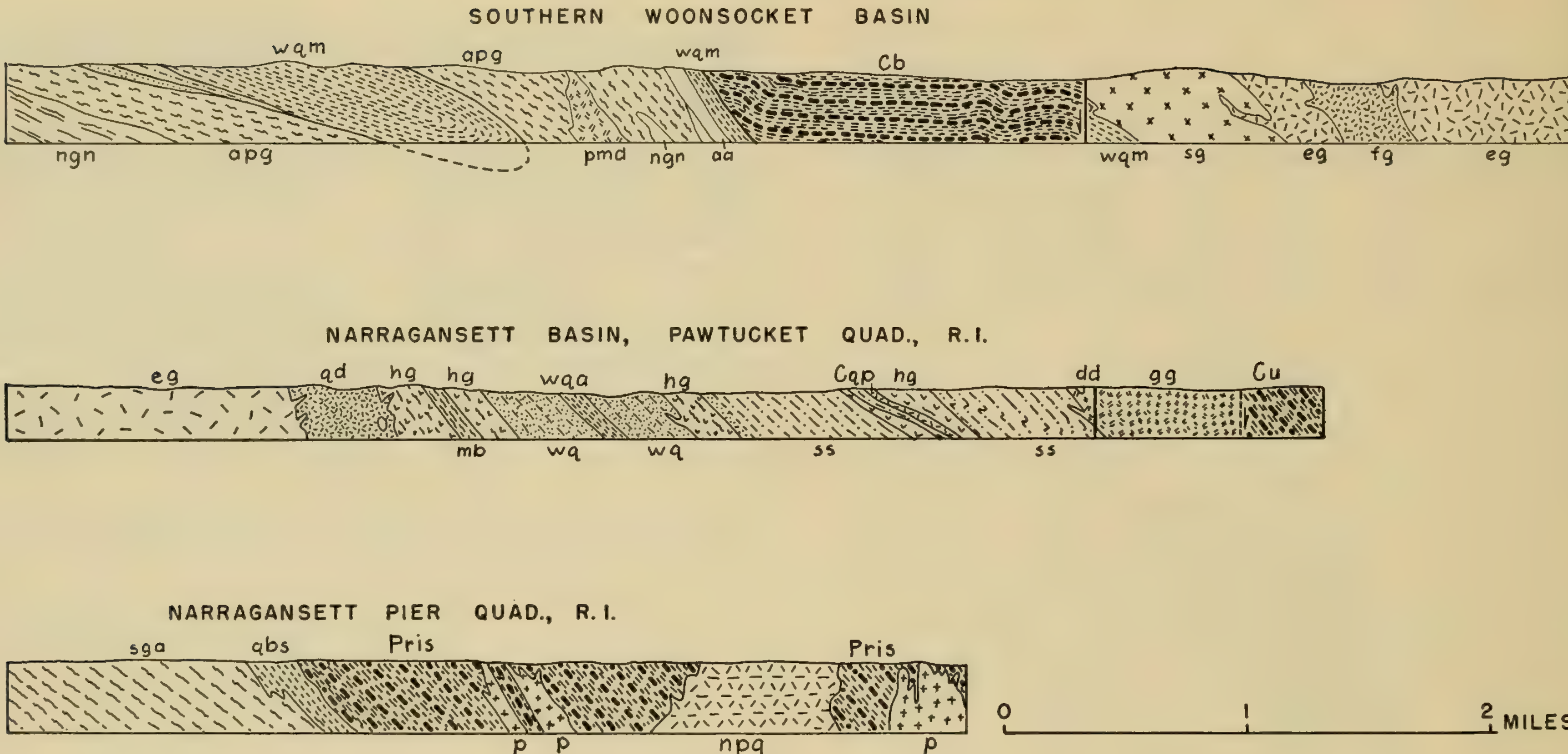


Fig. 11.30. Cross sections of Woonsocket and Narragansett basins. Top section after Richmond, 1952. ngn, Nipsachuck gneiss; apg, Absalona fm.; wqm, Woonsasquatucket fm.; pmd, metadiorite; Sg, Scituate granite gneiss; eg, Esmond granite; fg, fine-grained granite; Pb, Bellingham conglomerate.

Middle section after Quinn *et al.*, 1949. mb, Mussey Brook schist; wa, Westboro quartzite; wqa, Albion schist member; ss, Sneech Pond schist; hg, Hunting Hill greenstone; gg, Grant Mills granodiorite; Cqp, Quiney granite; Cu, Carboniferous undifferentiated; dd, diabase dike.

Lower section after Nichols, 1956. sgg, Scituate granite gneiss; qbs, Blackstone quartz-biotite schist; Pris, Rhode Island formation; npg, Narragansett Pier granite; p, pegmatite.

bles of the Esmond granite, their metamorphism was later than that of the Blackstone series. The later intrusives rocks of the East Greenwich group are essentially unmetamorphosed. The Pennsylvanian rocks are folded and fault-tilted, and schistosity is widespread. It is commonly not parallel to the bedding, and chloritoid, garnet, amphibole, biotite, and muscovite are developed.

In the Narragansett Pier quadrangle a reddish, massive to gneissic granite is clearly intrusive into the Pennsylvanian beds. It has been named the Narragansett Pier granite by Nichols (1956). A cross section is shown in Fig. 11.30. Elsewhere granites intrusive into the Pennsylvanian beds have been reported but the modern mapping casts doubt on such relations.

Woonsocket Basins

A section across the southern of the two small basins, here called the Woonsocket, is given in Fig. 11.30. The western margin of the Pennsylvanian basin dips steeply, although it is a sedimentary contact. The east margin is a high-angle normal fault contact (Richmond, 1952). The Bellingham conglomerate which fills the small basins generally dips eastward although it has many small and closely spaced folds. The west margin is a sedimentary overlap. The conglomerate pebbles are stretched in the plane of schistosity and the long axes point down dip. The matrix in places is a mica or chlorite schist which tends to enwrap the pebbles. The conglomerate in the southern basin is more sandy and less metamorphosed, and contains beds of graywacke, biotite-sericite schist, dark phyllite, and slate.

Boston Basin

The strata of the Boston basin comprise the Roxbury conglomerate below, and the Cambridge slate or argillite above. The Roxbury lies unconformably on the Dedham granodiorite of Precambrian (?) age, and is possibly Pennsylvanian and probably Permian in age, according to Billings *et al.* (1939). The conglomerate is over 3500 feet thick, and the slate about 3500 feet; both constitute the Boston Bay group. Part of the Roxbury conglomerate is volcanic and part sedimentary. The volcanic rocks include not only effusive lavas but also thick beds of tuff, agglomerate, volcanic breccia, and conglomerate.

The Roxbury conglomerate above most of the volcanics is described by Emerson as consisting of the Brookline conglomerate at the base, the Dorchester slate in the middle, and the Squantum tillite at the top. According to La Forge the threefold division does not persist throughout the area occupied by the formation with sufficient definiteness to warrant mapping the members separately. In some areas, beds like the Dorchester slate are intercalated in most of the formation below the tillite. The Brookline conglomerate is massive, coarse, and in some areas 1200 feet thick. It contains cobbles and boulders, many of which are of the underlying Dedham granodiorite or of the volcanic complex. The slate member is red and purple, and in one place possibly 2000 feet thick. Much of

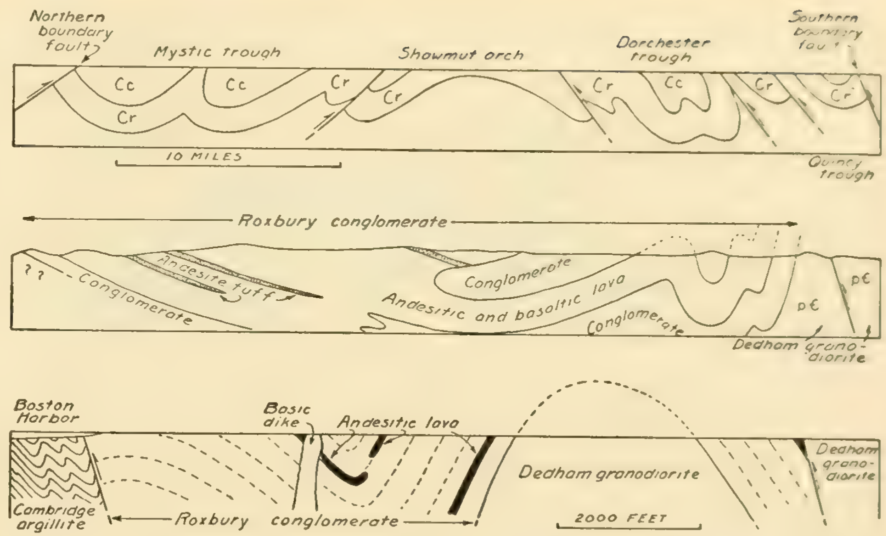


Fig. 11.31. Cross sections of Boston basin. Upper section from northwest to southeast across entire basin. Cr, Roxbury conglomerate; Cc, Cambridge slate; blank, pre-Carboniferous, mainly igneous (LaForge, 1932).
Middle section across Nantasket area. Section about 4000 feet in length. After Billings, Loomis, and Stewart, 1939.
Lower section across the Hingham area. After Billings *et al.*, 1939.

it is reworked basaltic and andesitic tuff, and layers of purple sandstone and grit are common. The Squantum tillite is exposed in many places in the southern part of the Boston basin, and is about 600 feet thick. It possesses many characteristics of glacial drift and is generally believed to have been deposited by local mountain glaciers.

The various lithologic types of the Roxbury conglomerate interfinger in a complex fashion in the Nantasket area, according to Billings (1939), and the formation consists of numerous lenses of sedimentary and volcanic materials overlapping one another. See cross sections, Fig. 11.31.

The Cambridge slate, over the Roxbury conglomerate, underlies nearly all the northern part of the Boston basin and occupies several long belts in the southern part. The rock is practically nowhere a true slate, but it generally has a dominant cleavage parallel with the bedding. It has vari-

Fig. 11.32. Gulf of Maine and continental shelf off Nova Scotia showing location of seismic profiles and Triassic basin in Bay of Fundy.

ously been called a pelite, shale, argillite, and slate. It contains some quartzite beds.

Dott (1961) believes the Boston Bay group may be mid-Paleozoic and not Carboniferous, and also that the Squantum is not a tillite but rather an orogenic clastic interfingered in the other lithologies.

Gulf of Maine

Cenozoic and Cretaceous Geology. The continental shelf extends eastward from Nantucket and Cape Cod, and a broad peninsula-like platform under less than 500 feet of water, bounded on the north by the Gulf of Maine and on the south by the deep Atlantic, is known as Georges Bank (Fig. 11.32). The Atlantic margin of the bank is trenched by deep submarine canyons, and from their walls have been dredged rock samples carrying both Tertiary and Upper Cretaceous fossils. Fragments of a coarse sandstone, Lower Monmouth or Upper Matawan (both Upper

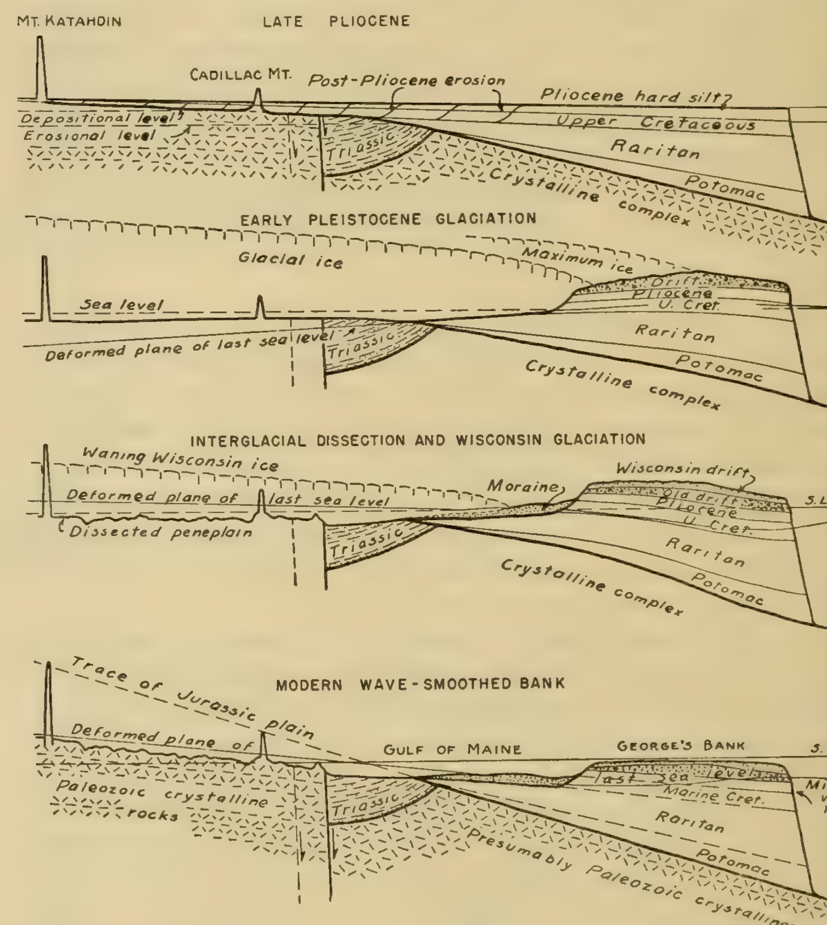


Fig. 11.33. Evolution of the Gulf of Maine and Georges Bank, generalized after a chart exhibited at the Geological Society of America meetings, 1948, by G. H. Chadwick and with his permission. Vertical scale greatly exaggerated.

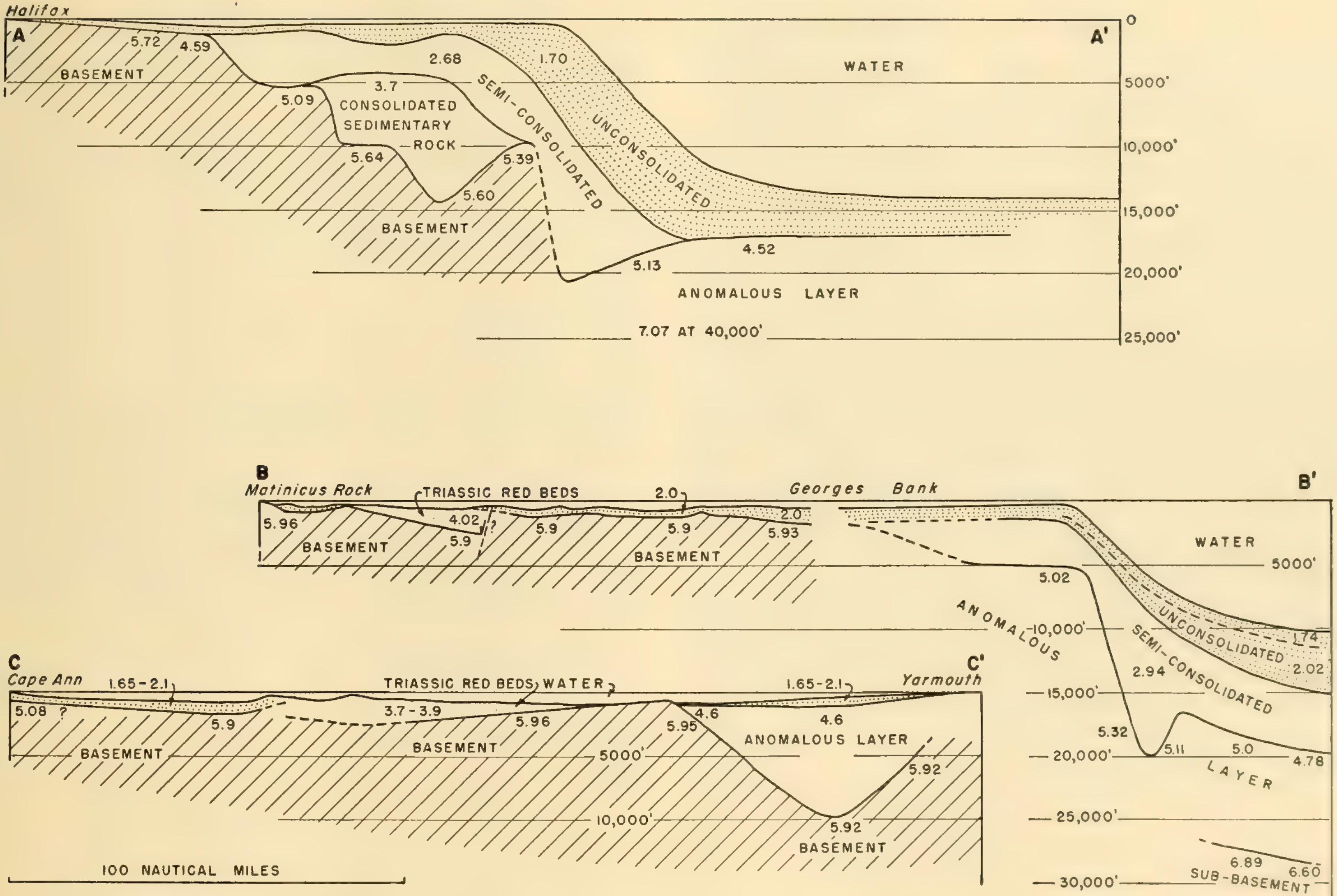


Fig. 11.34. Seismic profiles of Gulf of Maine and continental shelf off Nova Scotia. See Fig. 11.32 for location of profiles. After Drake *et al.*, 1954, and Officer and Ewing, 1954.

Cretaceous); of a glauconitic greensand, Navarro (equivalent of Monmouth); of an indurated green silt not older than Miocene; and of an impure glauconitic sandstone, late Tertiary in age, were broken from the walls of newly charted canyons cutting the southern margin of Georges Bank (Stetson, 1936). The thickness of the Tertiary sediments cannot exceed 1500 feet, and the top of the Upper Cretaceous ranges between 1450 and 1800 feet below sea level. Glacial drift and recent material mantle the gentler slopes, but in several places the older formations crop out on the steeper slopes. It is clear, therefore, that the Atlantic Coastal Plain, made up of Tertiary and Cretaceous sediments, continues eastward from the New York region and forms Georges Bank.

George H. Chadwick has prepared sections across the Gulf of Maine and Georges Bank showing the composition and evolution of the submerged coastal plain, and he has given permission to reproduce them, although they have not been published. See Fig. 11.33. These sections integrate the erosional surfaces, the sediments of Georges Bank as recognized by Stetson, an extension of the Nova Scotian Triassic trough, eustatic changes in sea level, and two stages of glaciation. The Lower Cretaceous Potomac is a projection from the New Jersey coastal plain and has not been sampled by the dredge.

Seismic Profiles. Refraction profiles have been run by Drake *et al.* (1954) and Officer and Ewing (1954) of the Gulf of Maine and continental shelf off Nova Scotia. These support the conclusions of Chadwick and Stetson as far as the unconsolidated and semiconsolidated sediments go (Cenozoic and Cretaceous). Compare Figs. 11.32 and 11.34, B–B′. It will be seen that the unconsolidated sediments are very thin over the Gulf of Maine and only thicken under the shelf slope.

The Triassic trough sediments are believed to exist, as Chadwick pictured them, on the basis of the layer that yielded the 3.7–4.02-km/sec velocities (Drake *et al.*, 1954).

The crystalline basement appears complicated by layers with lower than normal velocities. The 4.6-km/sec velocity layer south of Yarmouth (section C–C′), the 4.52–5.13-km/sec layer under the shelp slope and rise off Nova Scotia (sectionA–A′), and the 5.11–4.78-km/sec layer in the same place off Georges Bank (section B–B′) are the cases in point. They have been interpreted by Drake *et al.* to be part of the crystalline basement on the grounds that Katz *et al.* (1953) found two layers in Maine, recording at Falmouth, with a velocity of 5.34 km/sec for the upper and 6.24 km/sec for the lower. These are both somewhat higher than the presumed equivalents under the Gulf of Maine. In a study of the Outer Ridge and Blake–Bahama basin (reviewed in Chapter 10) a 5.2-km/sec velocity layer on a ±6.5-km/sec velocity layer was theorized to be a mass of extruded basalt on the typical “oceanic basalt” layer. The Gulf of Maine “anomalous layer” has velocities somewhat slower than the ″volcanic” layer under the Outer Ridge, and also lies on the crystalline basement—not on the ocean basalt layer. It would appear, therefore, that the anomalous layer is part of the Paleozoic complex of New England. It could be a mildly metamorphosed Carboniferous basin type of deposit, or conceivably a Mississippian (?) volcanic accumulation.

The floor of the continental shelf and shelf slope sediments off Nova Scotia and Georges Bank show a depression or trench similar to that off New Jersey. Refer to Fig. 10.6.

12.

MARITIME APPALACHIANS

DEFINITION

The Maritime Appalachians will here include the Paleozoic mountain systems of Nova Scotia, New Brunswick, Prince Edward Island, and the continuation of the structural elements of New York, Vermont, and New Hampshire in Quebec. The folded and thrust-faulted chains south of the St. Lawrence River extend northeastward into the Gaspé Peninsula, and all are intrinsically part of the Maritime geologic province. See index map of Fig. 11.1. The Maritime Appalachians, as here defined, are also known as the Appalachian-Acadian region (Alcock, 1947) and together with New England and Newfoundland, as Greater Acadia (Schuchert and Dunbar, 1934).

GEOMORPHIC PROVINCES

General Characteristics

The Martime Appalachians are made up of dissected uplands and broad lowlands. The shoreline is notably long and irregular, with many deep embayments. It is a fine example of a ria coast in which the linear structural elements run out under the sea. Figure 12.1 shows the physical divisions of New Brunswick and Nova Scotia which correspond to the following descriptions by Alcock (1947).

Nova Scotia

Nova Scotia is made up of five upland and as many lowland areas. The former comprise: (1) the large Southern Upland, which embraces the southern and central part of the peninsula and slopes from elevations of about 600 feet southeastward towards the Atlantic Ocean and also southwestward towards the Gulf of Maine; (2) North Mountain, a narrow, flat-topped belt, averaging about 550 feet high, that extends along the southeast side of the Bay of Fundy from Cape Blomidon in Minas Basin southwest for 120 miles to Brier Island; (3) the Cobequid Mountains, lying north of Minas Basin and stretching for 75 miles across Cumberland County from the head of the Bay of Fundy almost to Northumberland Strait; this region shows broad, rounded summits blending to form a somewhat rolling surface with an average elevation of a little more than 900 feet; (4) the highlands of eastern Pictou and Antigonish counties between New Glasgow and Antigonish and stretching northeastward to Cape George; in the southern part the average elevation is about 800 feet, but near Arisaig it is more nearly 900 feet; (5) the upland belts and northern tableland is the largest of these areas and presents an even flat-topped surface about 1,200 feet high.

The lowlands are underlain by less resistant rocks, such as sandstone, shales, limestone, and gypsum and show a considerable diversity of elevation and form. They comprise: (6) the Annapolis Cornwallis Valley, a long trough-like depression lying between the steep, straight wall of North Mountain and Colchester counties surrounding Minas Basin on the north, east, and south, and merging into Cornwallis Valley on the west; (7) the lowlands of Hants and Colchester counties surrounding Minas Basin on the north, east, and south, and merging into Cornwallis Valley on the west; (8) the Cumberland-Pictou area occupying all that part of the isthmus of Chignecto lying north and east of Cobequid Mountains; (9) the lowland of Antigonish and Guysborough counties, which lies south and east of the highlands extending towards Cape George; and (10) the lowlands of Cape Breton Island, areas lying between the upland belts and occupied by undulating country of landlocked lakes.

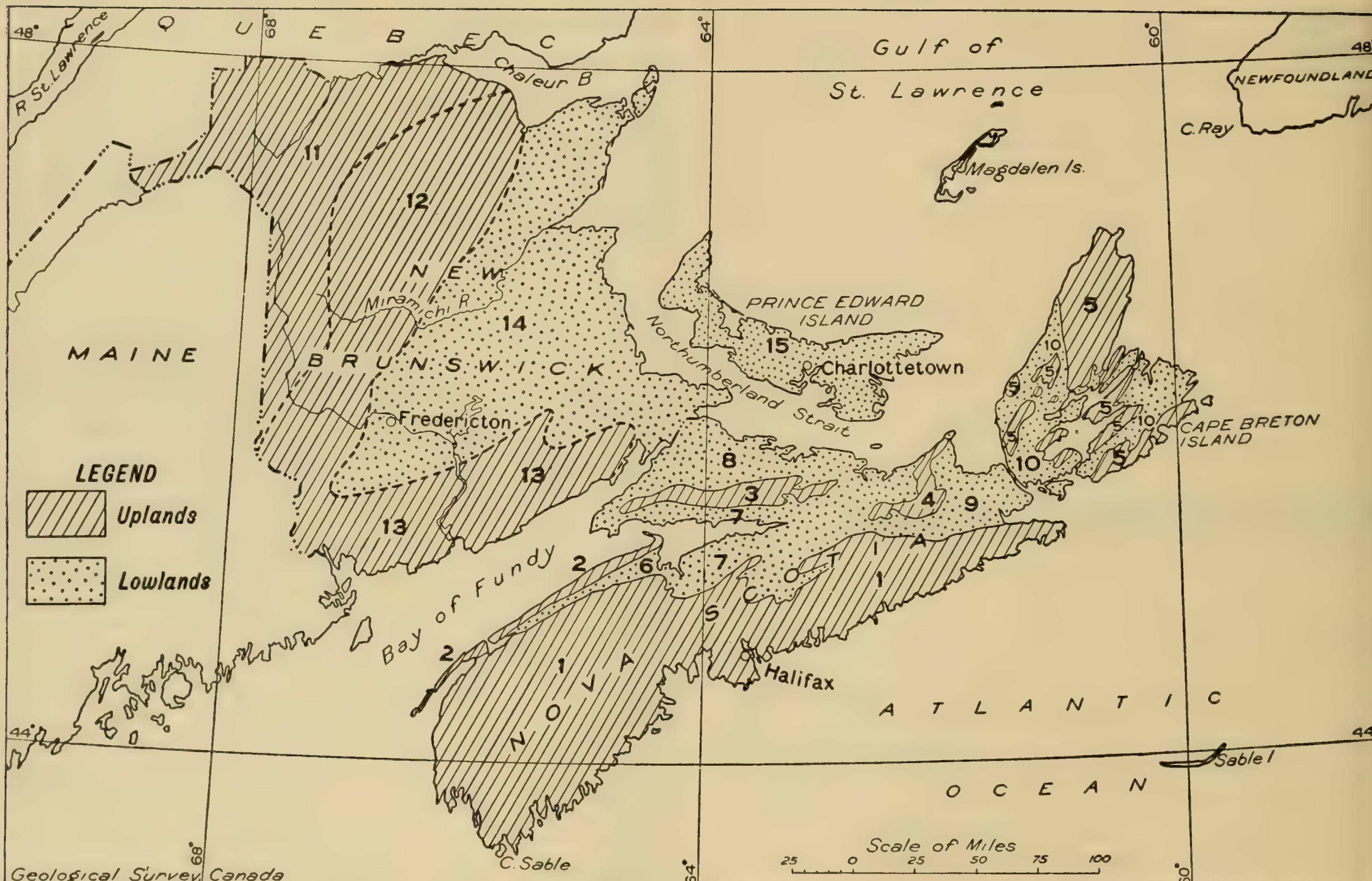


Fig. 12.1. Physical divisions of the Maritime Provinces, New Brunswick, Nova Scotia, and Prince Edward Island. Reproduced from Alcock, 1947.

New Brunswick and Prince Edward Island

New Brunswick falls naturally into four major topographic divisions whose boundaries, however, in most places are not sharply defined. The first, which may be regarded as the main axis of the province, is known as the Central Highlands, an upland region developed largely on resistant granitic, volcanic, and metamorphic rocks. It trends northeast through the central part of the province and is made up of ridges and hills, most of which have flat summits. Its elevation varies considerably, but much of it has an average height of about 1,000 feet. The highest part is where the tributaries of Miramichi, Nipisiguit, and Tobique Rivers take their rise. Here broad summits have a general elevation of about 2,200 feet, with some ridges and peaks rising to still greater heights. For example, Mount Carleton, the highest point in the province, has an elevation of 2,690 feet.

To the northwest of the Central Highlands is a second division, which may be termed the Northern Upland. It stands at an elevation of 800 to 1,000 feet above sea level and is developed on folded Paleozoic strata. The upland presents a remarkably uniform, flat-topped surface whose regularity is broken only by a few peaks and ridges rising slightly above the general level and by valleys such as those of the St. John and the Restigouche, which are deeply entrenched in it. The Stewart highway from Campbellton to St. Leonard crosses this belt.

The third division, the Eastern Plain, lies to the east of the Central Highlands, and makes up almost one-half of the province. It is a region of low relief, rarely more than 600 feet high, sloping gently to the Gulf of St. Lawrence. Its underlying rocks are mostly flat or gently dipping Carboniferous sediments. Prince Edward Island may be regarded as an outlier of this division, and the Cumberland-Pictou lowland area of Nova Scotia is continuous with it.

The fourth division, termed the Southern Highlands, lies along the Bay of Fundy. It is mainly an upland belt of ridges of which the most important is the flat-topped Caledonis Mountain belt of Albert, Kings, and St. John counties, which reaches a maximum elevation of 1,350 feet southeast of Markhamville. To the southwest, in Charlotte county, the belt merges into the Central Highlands. The region shows considerable topographic diversity and a great variety of rock types. The ridges are composed mainly of hard volcanic and intrusive rocks, whereas minor lowland areas within the belt have been carved from weaker strata.

Quebec

In Quebec the Appalachian region is bordered on the northwest by the St. Lawrence Lowlands into which it merges imperceptibly. In fact, considered from the point of view of topography, the lowland belt overlaps the Appalachian geological region. To the southwest the upland region includes three parallel groups of ridges and isolated hills and mountains. These are highest in the south, and decrease in elevation towards the northeast. The highest point is Round Top on Sutton Mountain, elevation 3,175 feet, near the Vermont border.

The most easterly of the three belts is known as the Megantic anticline. It forms part of the International Boundary, and to the northeast passes into Maine. To the west the Stoke Mountain anticline extends as far as Lake St. Francis, where it loses its identity. Still farther west, a little beyond Lake Memphremagog, the third range, the Sutton Mountain anticline, is a continuation of the Green Mountains of Vermont. Between the anticlinal ranges the country varies from 900 to 1,000 feet in elevation, presenting in places a remarkably level surface. To the northeast, it continues as an upland belt of ridges and rolling country cut across by deep valleys such as those of the St. Francis and Chaudiere. It decreases in elevation to a point about opposite Quebec City, but farther northeast it rises again and in the central part of the Gaspé Peninsula becomes the Shickshock Mountains, with elevations ranging up to more than 4,200 feet. The individual members of this range show broad flat summits and the range is bordered both to the north and south by another flat-topped upland at a lower level into which the present river valleys are deeply incised. On the north side of the Shickshock the descent to the lower upland is for the most part abrupt; on the south it is more gradual. The lower surface slopes off both to the north and to the south, and to the southwest merges with the Northern Upland of New Brunswick.

STRATIGRAPHY

Introduction

The Maritime Appalachians are a continuation of the New England Appalachians and present much the same geology. See geologic map of Fig. 12.2. They are composed mostly of Paleozoic rocks, both sedimentary and igneous, but some older Precambrian and some younger Triassic rocks are also present. The chart of Fig. 12.3 correlates the principal formations of Nova Scotia, New Brunswick, and Quebec, and may be referred to in the following brief enumeration of the stratigraphic systems. Several groups such as the Green Head, the George River, and the Coldbrook are known definitely to be Precambrian, and others such as the Meguma and Macquereau are regarded as Precambrian but on less satisfactory evidence. They may be Cambrian. Certain granite intrusions of the southern highlands of New Brunswick and in Cape Breton Island are also probably Precambrian, but absolute proof of this has not been established. Other belts of rock shown on early maps as Precambrian are now either definitely known or else inferred to be of Paleozoic age (Alcock, 1947).

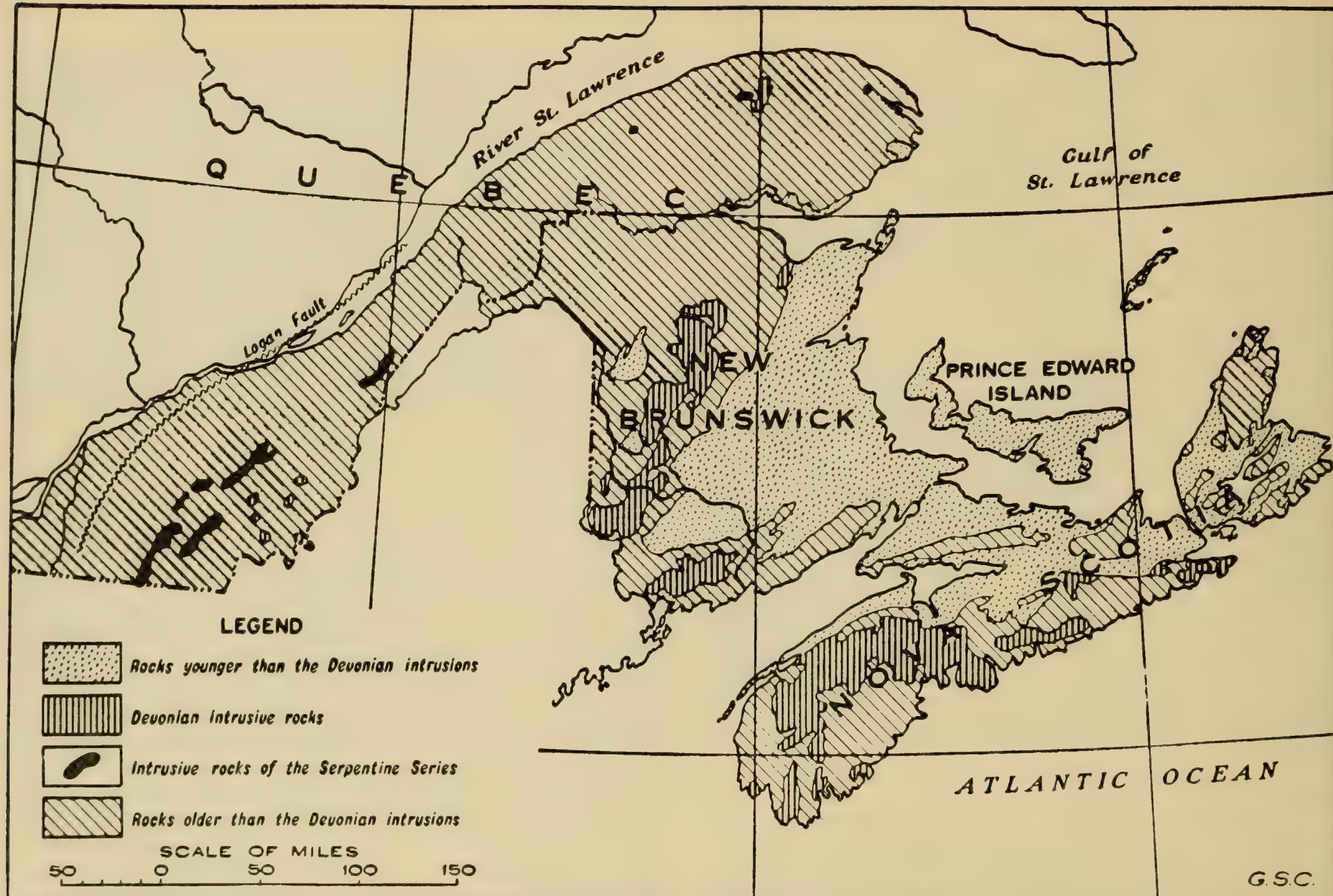


Fig. 12.2. Geologic map of the Maritime Provinces and Quebec. Reproduced from Alcock, 1947.

Cambrian System

Alcock (1947) reports that Cambrian rocks are found in southeastern Quebec, in Gaspé Peninsula, in southern New Brunswick, and in Cape Breton Island, Nova Scotia. In southeastern Quebec most of the rocks of this age are metamorphosed to a greater or less degree, and some are highly schistose. In the Oak Hill region near the Vermont border a series of Lower Cambrian strata 3000 to 4000 feet thick consist of slate, quartzite, dolomite, graywacke, and sericite schist. Rocks presumably of Cambrian age of the Thetford-Beauceville region, known as the Caldwell group, consist of nearly pure quartzites, slates, and pillow lavas of basaltic composition. A Cambrian seaway and trough of deposition probably extended from the Lake Champlain region to Quebec City and hence to Gaspé where some hard, gray limestone and ribboned, shaly limestone of late Cambrian age occur.

At St. John, southern New Brunswick, strata from Lower Cambrian to Lower Ordovician crop out, and these are known collectively as the St. John group. It consists of quartzites, limestones, and black shales. Similar beds occur on Cape Breton Island. They range in age from Lower to Upper Cambrian and consist of gray and black shales and slates with some quartzite and conglomerate, red sandstone and red and gray argillite carrying hematite, and greenish gray and reddish gray argillites.

Ordovician System

According to Alcock (1947):

In the Appalachian belt of Quebec, strata of Lower, Middle, and Upper Ordovician age are known, but in most places fossils are not sufficiently well preserved to permit an exact age determination. In the long belt from the Vermont border to the east end of Gaspé the deformed Ordovician strata were formerly referred to as the "Quebec group." This term had first been applied by Logan in 1860 to beds at Quebec City that had been thrust against and over the younger strata of Middle Ordovician age. Later the term became a convenient one to include all those early rocks whose exact age was unknown.

In Nova Scotia, Ordovician rocks are known to occur in the Pictou-Antigonish upland. They comprise metamorphosed sedimentary, volcanic, and intrusive varieties. The Browns Mountain group, consisting of argillites, slates and graywacke, is regarded on the evidence of a few fossil linguloids, as of Lower Ordovician age. Locally associated with the sediments are interbedded volcanic flows and tuffs, and cutting them is a stock of granite and dykes and stocks of rhyolite and quartz porphyry. In the Arisaig region, strata of this group are overlain by coarse conglomerate, and grit of the Malignant Cove formation, which is believed to be of Middle Ordovician age. In the Pictou region purplish red, arkosic conglomerate, purplish gray, arkosic grit, and purplish red argillite form what is known as the Stewart Brook formation, which is probably correlative with the Malignant Cove.

Era	Period	Epoch	Nova Scotia	New Brunswick	Quebec
Mesozoic	Triassic		Annapolis	Quaco; Lepreau	
	Permian				
	Carboniferous	Pennsylvania	Pictou; Morien; Stellarton Cumberland Riversdale	Clifton Lancaster Mispek } Petitcodiac	Bonaventure
		Mississipian	Canso Windsor Horton	Hopewell Windsor Moncton Albert Memramcook	
Paleozoic	Devonian	Upper Devonian		Perry	Escuminac Fleurant Pirate Cove
		Middle Devonian		Gaspe	Gaspe; Malbaie; Heppel
		Lower Devonian	McAdam Lake; Torbrook; Knoydart	Dalhousie	Grand Grève Bon Ami St. Albans; Dalhousie; Lake Aylmer
	Silurian		Arisaig; Kentville	Chaleur Bay; Mascarene	Chaleur Bay
	Ordovician	Upper Ordovician		Matapedia	Matapedia; Pabos; White Head
		Middle Ordovician	Malignant Cove; Stewart Brook	Tetagouche	Pohenagamooke; Mictaw; Quebec City; Beauceville; Farnham; St. Francis
		Lower Ordovician	Browns Mountain		Levis Sillery
	Cambrian		Boisdale	Saint John	Murphy Creek; Caldwell; Sutton; L'Islet
Proterozoic			Meguma (Gold-bearing)	Coldbrook	Macquereau; Tibbit Hill
Archean			George River	Green Head	

Fig. 12.3. Correlation chart of the principal formations of Nova Scotia, New Brunswick, and Quebec. Reproduced from Alcock, 1947.

In New Brunswick, rocks of Middle Ordovician age occur near Bathurst. Stretching to the southwest is a wide belt of sedimentary rocks, with, in places,

associated volcanic varieties. Much of this complex may be of Ordovician age. In the southwestern part of the province the Charlotte group is probably of Ordovician age. It is made up of two divisions, one known as the Dark Argillite, the other as the Pale Argillite. The former lies unconformably below strata of Silurian age and is composed of argillite, slate, quartzite, mica schist, gneiss, and minor amounts of volcanic rocks. It is intruded by masses of granite and gabbro. The Pale Argillite consists of argillite, sandstone, arkose, slate, and mica schist. In the St. Stephen area the beds are apparently comformable with and grade into those of Dark Argillite. On early maps the Pale Argillite was classed as Devonian on account of the reported finding on Cox Brook, a tributary of Magaguadavic River, of a Lepidodendron-like form. Later work has failed to find any fossils whatever in these rocks.

In the Thetford area, the Quebec group (Sillery and Levis) consists of black slates with a basal conglomerate and some interbedded impure quartzite or graywacke, overlying unconformably the Cambrian Caldwell group. In the Beauceville region volcanic tuffs and flows are interbedded with the sediments, and in places the series is so altered that it is difficult to distinguish the volcanic from the sedimentary members. Still farther southwest, near Phillipsburg in the Lake Champlain region, a thick series of fossiliferous Beekmantown sediments consisting of shales and limestones overlies Upper Cambrian beds and is followed by strata of Chazy of Middle Ordovician age.

To the northeast of Levis, rocks consisting of red, green, gray, and black slates, quartzites, and conglomerates form a belt in places 20 miles wide. These beds have been correlated with the Sillery, but both younger and older strata may be included. An interesting feature in these rocks is the presences of belts of limestone conglomerates. These occur at various horizons in both the Sillery and the Levis, forming bands from about a foot to more than 100 feet in thickness. The pebbles and boulders consist of gray limestone, and weigh from less than an ounce to many tons. Similar limestone conglomerates are found in Newfoundland to the northeast and Vermont to the southwest. They have been interpreted as the result of local slipping and breaking up of limestone along the sea bottom by earthquakes in a zone where faulting was prevalent. Another feature of the Sillery is the occurrence of belts of quartzite, locally called the Kamouraska formation. These belts are lenticular but extensive, and their thickness varies greatly.

Interbedded arkose and volcanic rocks of Ordovician age are known in the Shickshock Mountains; and dark shales, limestones, conglomerates, argillites, quartzose sandstone, and volcanic flows and tuffs occur to the south on both sides of Chaleur Bay.

Silurian System

The best Silurian section in Nova Scotia is at Arisaig where 3800 feet of highly fossiliferous sandstones and shales occur. The series is overlain by Lower Devonian beds, and it overlies a flow of rhyolite probably of Lower Ordovician age. The faunas can be correlated better with British than with American; even the resemblances with the Chaleur Bay Silurian faunas are slight.

On the north side of Chaleur Bay is probably the thickest marine Middle Silurian succession in North America. At the top of the sequence are volcanic flows interbedded with sediments, chiefly clastics, and flows are present also in other formations farther down in the succession. A total of 8427 feet or more of sedimentary rocks and 4626 feet of volcanic rocks are present in the Black Cape area.

In southern New Brunswick, on the Bay of Fundy, great quantities of volcanic rocks, chiefly rhyolites and andesites, are interbedded with sediments. At Oak Bay a basal Silurian coarse conglomerate rests unconformably on the dark argillite of the Charlotte group of Ordovician age. The belt is a continuation of one extending from the Eastport area of Maine, where a number of formations of Middle and Upper Silurian age occur.

Devonian System

Rocks of Lower Devonian age occur in Quebec, New Brunswick, and Nova Scotia. Sedimentation at this time was accompanied by widespread volcanism, and at the close of the epoch the main phase of the Acadian orogeny took place. In the Middle Devonian, great thicknesses of clastic sediments accumulated in the Gaspé Peninsula, and in Upper Devonian time sedimentation progressed locally in the Chaleur Bay and Bay of Fundy regions (Alcock, 1947).

A well-known Lower Devonian section is at the eastern end of Gaspé Peninsula, where about 2000 feet of limestone and limy shale beds have been described. Within central Gaspé Lower Devonian shales and limestones, associated with thick deposits of volcanic rocks, are widespread. At the west end of the peninsula, shales and argillaceous limestones of the same age are 2200 feet thick.

The Lower Devonian rocks at Dalhousie consist of highly fossiliferous marine sediments, volcanic flows, and tuffs, dikes, and volcanic rocks. The principal Nova Scotian Lower Devonian section is southwest of Arisaig, where fine-grained, red, arenaceous slates and gray sandstones

1000 feet thick, and apparently of continental origin, overlie with marked erosional unconformity Silurian strata of the Arisaig series.

Much of the interior of Gaspé is underlain by sandstones, conglomerates, and arenaceous shales varying in color from green and drab to red. The type locality is on Gaspé Bay where a section 7000 feet thick rests on the Lower Devonian limestones.

Upper Devonian beds are present on the north side of Chaleur Bay in a three-unit sequence. The lower formation consists of pebble conglomerates and sandstones and 450 feet of coffee-colored shale. The middle formation consists of a coarse pebble-and-boulder conglomerate with gray matrix. It is only 45 feet thick. The upper formation is 385 feet thick and consists of gray shales and shaly sandstones.

On the western side of Passamaquoddy Bay, in the St. Andrews region of New Brunswick, near the Maine border, on the opposite side of the bay on Mascarene Peninsula, at Black Harbour south of St. George, and on some of the adjacent islands are areas underlain by beds of red sandstone and conglomerate that are correlated with the Perry conglomerate of Maine.

The beds lie for the most part with low dips and in gentle folds. In places they rest unconformably on the Silurian rocks, and in places are in fault contact against them. The conglomerates contain boulders of the Silurian and pre-Silurian rocks and of the St. George granite intrusive rocks. On Hill Island two basic amygdaloidal lava flows are interbedded with the red sediments, and similar volcanic rock shows on Howard Island. Locally the beds are cut by dark dykes. Similar dykes and flows are associated with the conglomerate beds at St. Andrews (Alcock, 1947).

Carboniferous System

Carboniferous strata underlie extensive areas of New Brunswick and Nova Scotia. They also underlie all of Prince Edward Island and the Magdalen Islands in the Gulf of St. Lawrence, and they crop out along the north shore of Chaleur Bay. They represent Mississippian, Pennsylvanian, and possibly part of Permian time, and are the source of coal, oil, gas, and gypsum in New Brunswick and Nova Scotia. They are generally softer and more susceptible to erosion than the older Paleozoics and form the lowlands. The lowlands of the geomorphic map of Fig. 12.1 are, therefore, about coincident with the Carboniferous beds. See also the *Geologic Map of North America.*

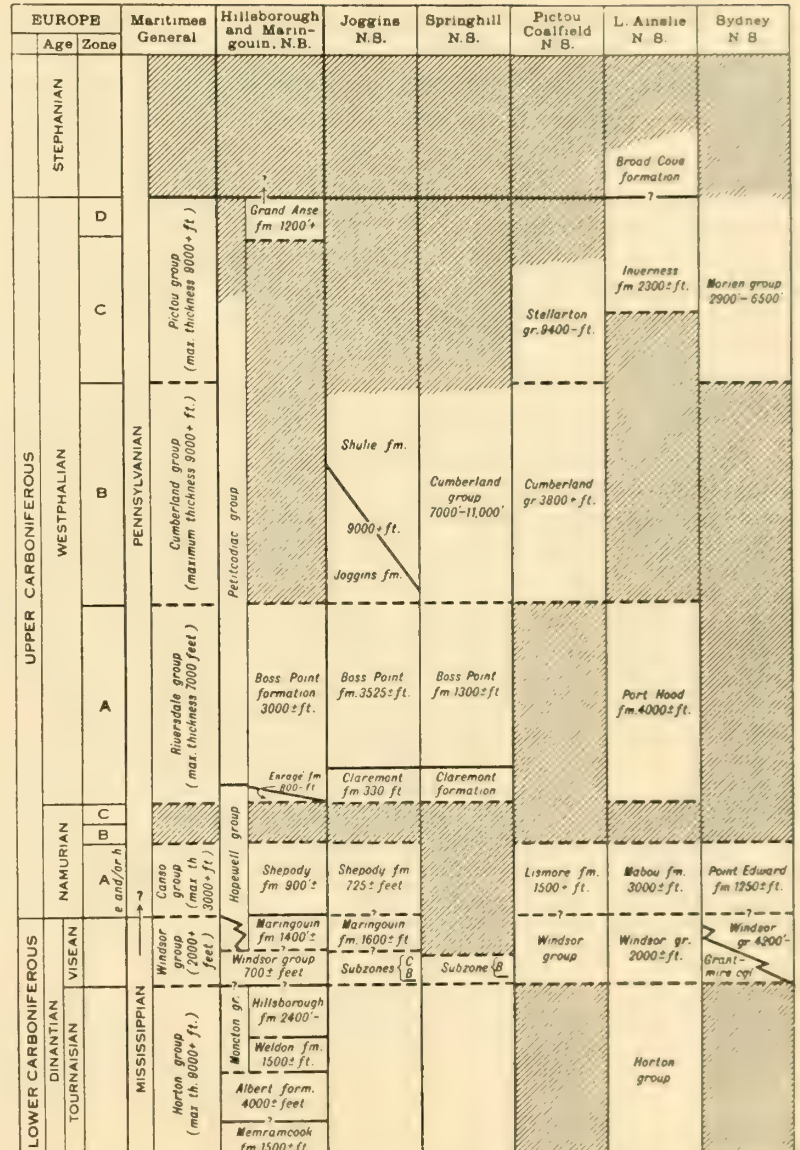


Fig. 12.4. Correlation chart of the Carboniferous formations of New Brunswick and Nova Scotia. Reproduced from Alcock, 1947.

The Carboniferous strata make up extremely thick sequences, are dominantly conglomerates, sandstones, and shales; they contain several angular unconformities, and are particularly instructive of crustal unrest and of the tectonic history of the region. The correlation chart of Fig. 12.4 gives the important formations of the Carboniferous rocks in the Maritime Provinces. From it some idea of the numerous units, large thicknesses, and unconformities can be gained. The sedimentary and tectonic history is even more detailed than the chart indicates. For instance, the Mississippian strata of Nova Scotia belong to two groups, the Horton and the Windsor; and along the lower part of the Avon River, the Horton group

> . . . is made up of two formations, a lower known as the Horton Bluff, consisting of some 3,400 feet of dark shale, sandstone, and conglomerate, and an upper, the Cheverie, made up of 600 feet of red and grey shales, sandstone and arkose. The Horton Bluff formation rests unconformably on pre-Carboniferous metamorphic and igneous rocks; it contains plant remains, buried forests, and soils, and has a fauna of ostracod, crustaceans, and fish remains. The Cheverie rests with an angular unconformity on the Horton Bluff and is succeeded, also unconformably, by the Windsor group of marine sediments. The latter comprise limestone, gypsum, shale, sandstone and limestone conglomerate, the whole having a thickness of about 1,550 feet. The limestone members are rich in fossils and have yielded one hundred and twenty-seven species, chiefly molluscs and brachiopods.
>
> The Mississippian rocks extend eastward through the lowland belt to the Strait of Canso, and also occupy much of the lowlands of the southwestern part of Cape Breton Island. In the Lake Ainslie area, the Horton group includes about 6,000 feet of conformable, dominantly clastic sediments containing a meagre flora and fauna. They are intruded by diabase dykes and sills. The succeeding Windsor beds have here a thickness of about 2,000 feet. In the Arisaig region, diabase and basalt dykes and stocks intrude red conglomerate, sandstone, and sandy shale of the Mississippian McAras Brook formation, but are apparently older than the limestone of the Ardness formation of Mississippian age (Alcock, 1947).

The Pennsylvanian rocks of Nova Scotia are wholly nonmarine, as far as known, are dominantly clastic and red, and contain locally beds of coal and thin limestones. Pennsylvanian rocks also cover much of the plain of eastern New Brunswick, being made up of red and gray shales, sandstones, grits, and conglomerates.

> The north shore of Chaleur Bay is bordered for considerable distances by red clastic beds of the Bonaventure formation, which takes its name from the Bonaventure Island at Perce. The strata consists chiefly of coarse conglomerates and sandstones, with associated red shales, shaly sandstones, and locally limestone. The beds for the most part lie horizontally, but are locally tilted and in places faulted.

For relations along the north shore of Chaleur Bay see Fig. 12.5.

Magdalen Islands are composed of folded sedimentary and volcanic rocks of Mississippian age, surrounded by flat-lying beds of red sandstone of probable Pennsylvanian age.

Triassic System

Red sandstones, shales, and conglomerates of Triassic age occur in the Bay of Fundy region. They are most extensive on the southeast side of the bay, where a belt stretches along the entire length of the bay and borders both sides of Minas Basin. See Fig. 11.31. They rest unconformably on various Paleozoic and Precambrian formations and are capped by about 1000 feet of basaltic lavas that form the North Mountain upland. On the northwest side of the Bay of Fundy, patches of similar red conglomerate and sandstone occur. The beds of all these patches dip to the northwest and are in fault contact with the older formations. It has been concluded that they are deposits in a down-faulted basin similar to those of the Triassic red beds in the Connecticut and New Jersey basins. This is the northeasternmost of the known Triassic fault basins in the Appalachian mountain systems. It is believed to extend under the Gulf of Maine nearly to Boston. See Fig. 11.31.

IGNEOUS ROCKS

Extrusive Rocks

Interbedded volcanic rocks of various kinds have already been mentioned in the account of the stratigraphy. They are known in the Cambrian and Lower Ordovician of the Thetford–Beauceville region of Quebec, in the Middle Ordovician in the central Shickshock Mountains, and on the south side of Chaleur Bay. They are also known in the Middle Silurian in various places on the Gaspé Peninsula on the north side of Chaleur Bay, along the New Brunswick side of the Bay of Fundy, and

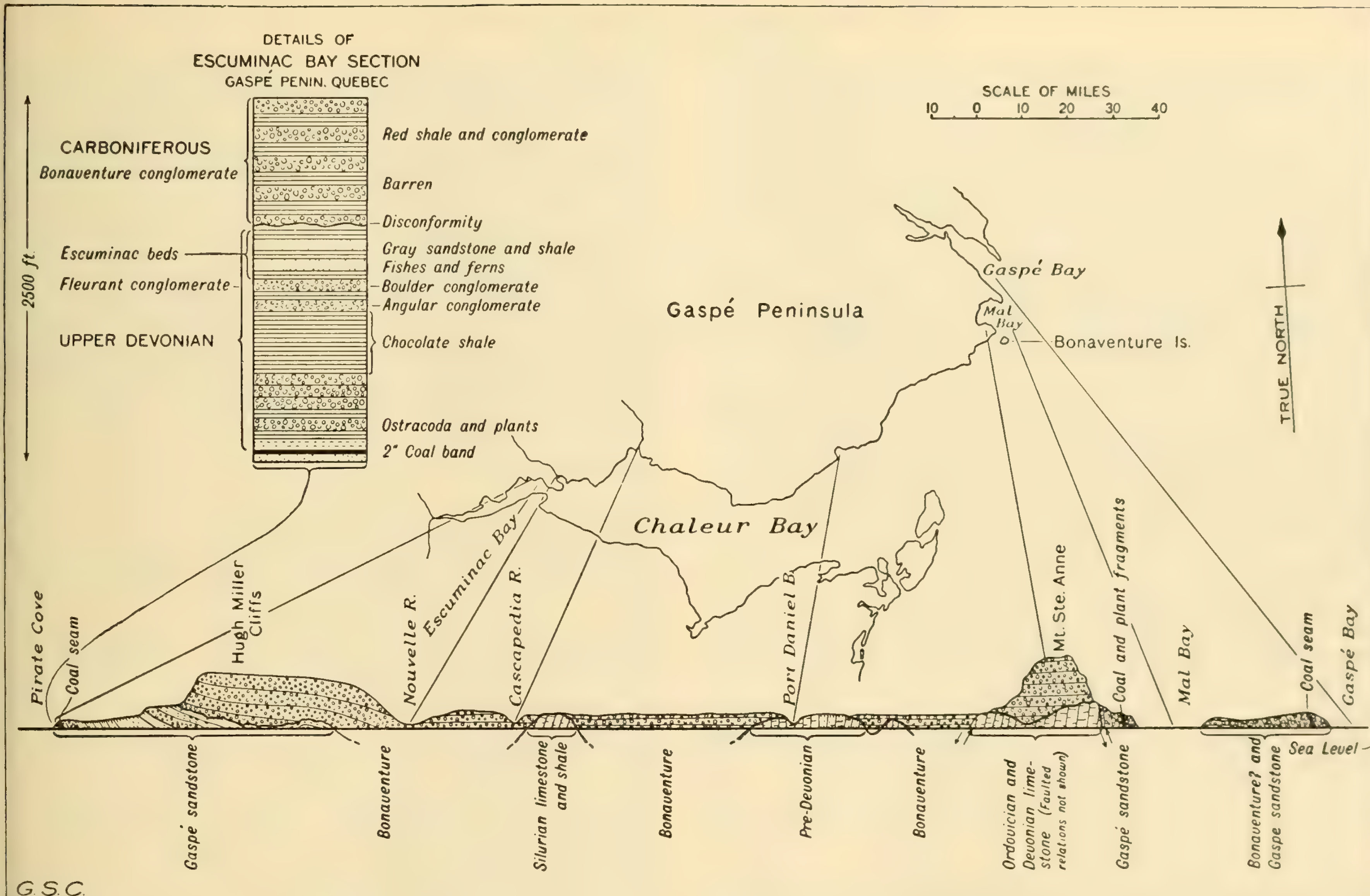


Fig. 12.5. Diagrammatic section along the north shore of Chaleur Bay. Reproduced from Alcock, 1947. Gaspé sandstone is middle Devonian.

in the Eastport area of Maine. The Silurian outpourings were especially voluminous and, where identified, are chiefly andesites and basalts, although acidic varieties have been noted. Volcanism was again widespread and voluminous in the Devonian. Lower Devonian volcanics are known in the Gaspé Peninsula, in northern New Brunswick, and in the Lake Ainslie area of Nova Scotia; Upper Devonian lavas have been noted in the St. Andrews region of New Brunswick near the Maine border. The Devonian volcanics are mostly andesites.

The Carboniferous was unremitting in volcanic activity, and considerable amounts of lavas and tuffs were extruded. The Mississippian rocks of the Magdalen Islands contain basaltic lavas and fragmentals, and those of the Hampstead area of New Brunswick contain rhyolite. Pennsylvanian rocks in the St. John region of the Bay of Fundy contain extrusive and intrusive rocks, and the Bonaventure formation along the north shore of Chaleur Bay contains amygdaloidal basalt flows.

Lavas, chiefly andesitic and basaltic, and graywackes and arkoses with sandstones, shales, and limestones compose a stratified assemblage typical of the eugeosyncline of Kay (1951).

Intrusive Rocks

Intrusive rocks are widespread in Nova Scotia and New Brunswick. They are granites and associated differentiates, that accompanied the Acadian orogeny at the close of Lower Devonian time. The granites are exposed over much of the southern upland of Nova Scotia, and the central highlands of New Brunswick.

A belt of ultrabasic plutons, now largely serpentinized, extends through the Quebec Appalachians from Vermont to Gaspé, and their intrusion is thought to have accompanied the Taconic orogeny. See Fig. 8.29.

Many dikes and sills are mentioned in the literature, and these probably relate to the volcanic series.

A group of eight small intrusions in southern Quebec form the Monterigian Hills. The most westerly is Mount Royal at Montreal. Except for one, they lie along a curved line that extends easterly from Montreal. Five of them rise well over 1000 feet above the surrounding plain; the others to heights of 600 to 700 feet. The five westerly ones intrude the flat-lying beds in front of Logan's line, and the three easterly ones cut the folded and faulted Paleozoics east of the line. According to Caley (1947):

> Brome and Shefford Mountains are thought to be unroofed laccoliths, or perhaps parts of a single laccolith still covered by sedimentary strata in the 2½ mile interval of lower land between the hills. The remaining hills appear to be volcanic necks with nearly vertical walls.
>
> The age of the intrusions is Devonian or younger. Evidence for this, in addition to that supplied by the St. Helen Island breccia, is afforded by Yamaska, Shefford, and Brome Mountains, which lie within the folded Appalachian region. The intrusive masses show no effects of deformation, and hence must have been intruded after the last folding that affected this region in Middle Devonian time. It has also been noted that in the Monterigian intrusive rocks pleochroic haloes surrounding crystals of zircon and titanite are invariably poorly developed and immature. In this they resemble those in Tertiary intrusive rocks, whereas in certain Devonian granites haloes are numerous and prominent. The suggestion has, therefore, been advanced that the igneous rocks of the Monterigian Hills may be as young as Tertiary.

STRUCTURES

Unconformities

The Paleozoic section is replete with unconformities and conglomerates which indicate intermittent orogeny from place to place over a long time.

A coarse conglomerate of Lower Cambrian age containing large granitic boulders rests on rocks of the same material as the boulders near St. John, New Brunswick. Lower Ordovician black slates with a basal conglomerate and some interbedded impure quartzite or graywacke overlie unconformably the Caldwell group of the Cambrian in the Thetford area of southern Quebec. Limestone conglomerates of Lower Ordovician age occur in places from Vermont through Quebec to Newfoundland and have been interpreted as the result of local slipping and breaking up of limestones, just deposited, along the sea bottom by earthquakes in a zone of crustal deformation.

In Nova Scotia, a coarse conglomerate and grit of Middle Ordovician age overlies beds of Lower Ordovician age. On the north side of Chaleur Bay, coarse conglomerates of Middle Ordovician age made up largely of the Proterozoic (?) Macquereau rocks, rest on the Macquereau. In the

same general area is a broad belt of Upper Ordovician conglomerate and grit about 2000 feet thick. Its relations to underlying beds are not noted.

The Arisaig Silurian series of Nova Scotia contains conglomerates and rests on Lower Ordovician volcanics. At Oak Bay in southern New Brunswick, the base of the Silurian succession is a coarse conglomerate which rests unconformably on the dark argillite of Ordovician age.

Lower Devonian red slates and gray sandstones southwest of Arisaig overlie with a marked unconformity Silurian strata. Arkoses and conglomerates occur in the Lower Devonian of Cape Breton Island. Much of the interior of the Gaspé Peninsula is underlain by Middle Devonian sandstones, conglomerates, and arenaceous shales. The change from limestones of the Lower Devonian to clastics of the Middle Devonian is generally regarded here as marking the principal phase of the Acadian orogeny. In the zinc and lead district of Berry Mountain and Brandy Brooks, the limestones are cut and mineralized by granitic and syenitic intrusive rocks, but not the overlying sandstones.

Upper Devonian beds on the north side of Chaleur Bay consist at the base of about 600 feet of coarse conglomerates and sandstone. These have been cast into a broad syncline, eroded, and are unconformably overlain by the Pennsylvanian Bonaventure conglomerate. More conglomerates of the Late Devonian age occur in New Brunswick near the Maine border; they are correlated with the Perry conglomerate of Maine. These beds are seen to rest unconformably on the Silurian rocks, and they contain boulders of the Silurian and pre-Silurian rocks of the St. George granitic intrusives.

The Carboniferous sediments rest everywhere, it is believed, in marked angular unconformity on older rocks, which range from Precambrian to Late Devonian in age. They are thousands of feet thick and contain great quantities of coarse clastics, particularly the Pennsylvanian. In Nova Scotia, the Horton Bluff clastics at the base of the Mississippian rest unconformably on pre-Carboniferous metamorphics and igneous rocks, and are in turn separated by an angular unconformity from the overlying Cheverie, also of Mississippian age.

Mississippian limestones and volcanics are folded, eroded, and overlain unconformably by Pennsylvanian (?) strata on the Magdalen Islands. Gussow's (1953) studies in New Brunswick lead to the conclusion that the Lower Mississippian strata rest unconformably on the older Acadian complex, and in turn are overlain unconformably by the Upper Mississippian strata. The Upper Mississippian strata were in turn strongly folded and faulted, eroded, and overlain unconformably by the Pennsylvanian clastics. The structure imposed on the Mississippian strata, both during and at the close of the period, is typically Appalachian-type folding and thrust faulting. The Pennsylvanian strata have not been disturbed to any extent since deposition and are essentially flat. The great amount of conglomerate attests the vigorous elevation of sizable highlands immediately before and during deposition.

Folds and Thrusts

All pre-Carboniferous rocks are considerably deformed and in part metamorphosed. In places, the Carboniferous strata are also deformed. The chief structures are folds. Some thrusts have been observed and mapped, particularly in New Brunswick, but for the most part mapping has not been sufficiently detailed to bring out the existence of long and great thrust sheets. The linear elements in the compressional structures trend generally northeastward in continuation of those of New England. The folds and thrusts of the Taconic and Acadian systems of New York and Vermont pass into southern Quebec, and the Taconic front reaches the St. Lawrence at Quebec City. There the Quebec formation carries Trenton fossils, and consists of limestone and shale and thin beds of limestone conglomerate. See Fig. 12.6. Its beds have been altered and cleaved. Beds of the older Levis formation have been thrust from the southeast against the Quebec City, whereas on the northeast the Quebec City is thrust against and over younger Upper Ordovician beds, the Utica-Lorraine. The Utica–Lorraine in turn is in contact with the Precambrian of the Canadian Shield. Resting horizontally and free from disturbance, directly on the Precambrian, are Trenton limestones unlike the beds of the Quebec City formation of similar age. This boundary between the highly deformed and the undeformed strata has long been known as Logan's line or Logan's fault (see map, Fig. 12.2). From Quebec City the line runs under the waters of the St. Lawrence, and sweeps in a

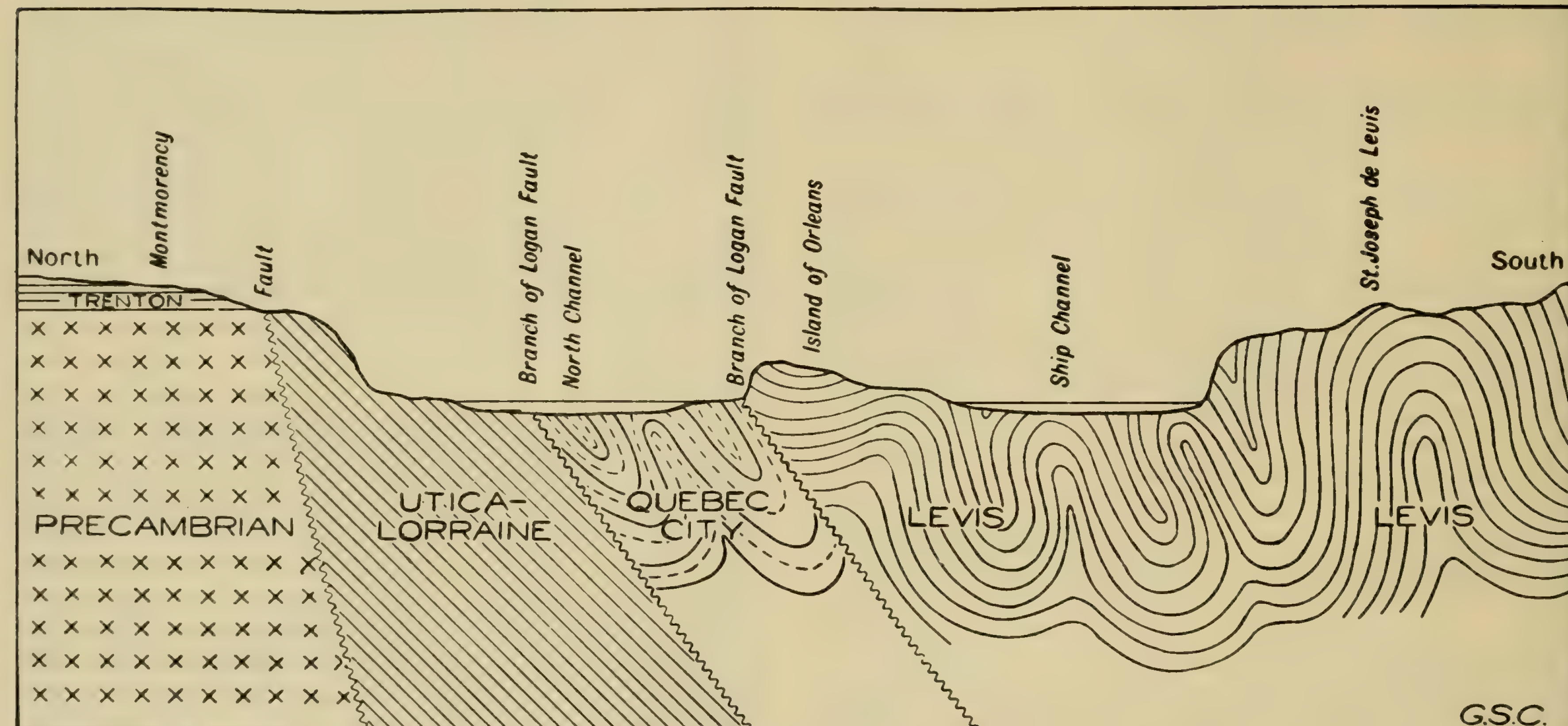


Fig. 12.6. Section across the St. Lawrence at Quebec City. Reproduced from Alcock, 1947.

smooth curve easterly to the tip of Gaspé, passing between Anticosti Island and the peninsula. Where information is available, the faults in the great arcuate zone of deformation south of the St. Lawrence are known to be overthrusts from the southeast. The rocks of this belt, particularly those of Gaspé, can be divided into four main assemblages according to the number of orogenies by which they have been affected. The metamorphic rocks of the Macquereau group were deformed by a late Proterozoic to early Cambrian orogeny; the Upper Cambrian and Ordovician rocks were deformed by the Taconic orogeny; the Silurian and Devonian rocks were affected by the Acadian orogeny; the Carboniferous is comparatively little disturbed (Alcock, 1935). Figure 12.5 illustrates the structures in a small way.

By reference to the *Geologic Map of Canada,* it will be seen that the lower and outer part of Nova Scotia is made up of Precambrian rock, as well as a belt through St. John, New Brunswick. These were not immune to Paleozoic orogeny, however, because overlying and marginal Paleozoic strata are much deformed and the Precambrian rocks are intruded by many plutons of Paleozoic age.

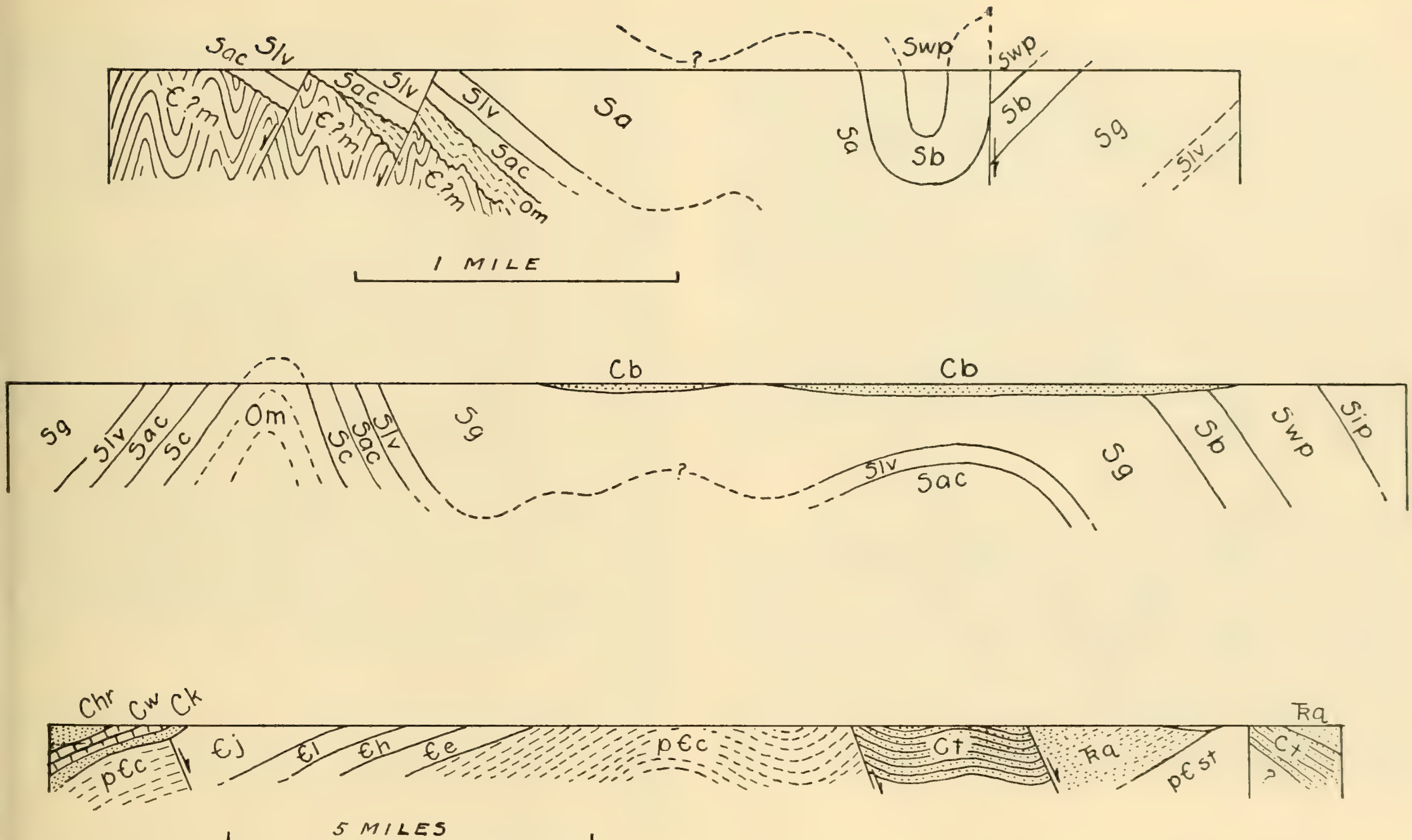


Fig. 12.7. Cross sections in the Maritime Provinces. Upper two sections are near Port Daniel Bay or the south coast of Gaspé Peninsula. After Northrop, 1939. Є?m, Macquerean metaclastics, either Cambrian or Precambrian; Om, Mictaw clastics; Sc, Clemville formation; Sac, Anse Cascon formation; Slv, LaVieille formation; Sg, Gascons formation; Sb, Bouleaux formation; Swp, West Point formation; Cb, Bonaventure formation. All Silurian formations are Middle Silurian in age. Lower section is of the St. John area, New Brunswick. After Hayes and Howell, 1937. pЄc, Cold Brook volcanics; pЄst, St. Martin volcanics, conglomerates and intrusives; Єe, Etcheminian sandstone; Єh, Hanfordian formation; Єl, Loch Lomond slate; Єj, Johanian sandstone; Ck, Kennebecasis conglomerate; Cw, Windsor limestone; Chr, Red Head conglomerate; Ct, Tynemouth Creek conglomerate; Trq, Quaco clastics.

The Arisaig region of Nova Scotia was affected by folding and intrusives at the close of Lower Ordovician and probably again at the close of the period, when the Taconic orogeny spread over much of the Maritime Provinces. Numerous plutons, mostly of Middle Devonian age, were emplaced in the Nova Scotian Precambrian and in the pre-Devonian strata of central New Brunswick as previously described. Similar intrusions occurred in the Gaspé Peninsula. The strata of New Brunswick and Nova Scotia were cast into northeasterly trending folds at this time, which probably paralleled former structures. Figure 12.7 shows the folds and faults of the St. John area in New Brunswick.

Normal faults are shown in a number of cross sections in the literature but are not much discussed. They are evidently later than the compressional orogenies or due to late adjustments of the individual orogenies. Some may be related to the Triassic basin faults and some even to Tertiary faulting.

TECTONIC HISTORY

Most writers emphasize two great orogenies in the Maritime Provinces, the Taconic at the close of the Ordovician and the Acadian or Shickshokian that ran its course through middle and late Devonian time. If the geologist is not influenced unduly by the interpretations and conclusions of numerous writers and considers only the numerous coarse conglomerates, unconformities, and volcanic series without previous impressions, it would be natural to conclude that a long succession of compressional impulses with accompanying intermittent volcanic and magmatic activity affected the Maritime Provinces. At intervals from Proterozoic to late Triassic time, vigorous deformation occurred from place to place. It does not seem altogether sound to the writer to conclude that two orogenies stand apart as clear-cut and distinct. Perhaps orogenic events reached maxima, and these maxima are to be considered the Taconic and Acadian orogenies. The great angular unconformity at the base of the Carboniferous emphasizes the superior nature of the orogenic phases that preceded the Mississippian.

The Mississippian beds are folded in places, and so are the Pennsylvanian, but the phases of Carboniferous orogeny are not so severe as the earlier ones. Over the New Brunswick lowlands the beds are fairly flat. Bordering highlands were intermittently and sharply elevated, however, throughout the Mississippian and Pennsylvanian to supply the great amounts of coarse clastics that make up the thick deposits. One of these source areas probably was the Precambrian area of Nova Scotia; another possibly lay to the northeast along the St. Lawrence.

During the succession of orogenies that beset the Maritime Provinces, several ranges were undoubtedly elevated and several troughs of deposition undoubtedly sank, and this activity was accompanied by voluminous volcanism. With the sea extensively invading the cordillera, a condition is visualized much like the partially submerged Andean system of southern Chile, Patagonia, and Tierra del Fuego. The changing geographic scene during the Paleozoic has not been set down on maps—perhaps the geological information is not yet sufficient to perform such a snythesis.

The fronts of the orogenic belts, however, seem clear by now, and after the geology of Newfoundland has been presented, an attempt will be made to relate the orogenic belts of Greater Acadia.

13.

NEWFOUNDLAND APPALACHIANS

Fig. 13.1. Physical divisions of Newfoundland. After A. K. Snelgrove, Newfoundland Geological Survey. The horizontally ruled areas are upland and the unruled areas, lowland. Small numbered uplands are 1, Hare Bay serpentine hills; 2, Highlands of St. John; 3, Indian Head Range. The lowlands take appropriate names such as West Coast Lowland; Grand Lake–White Bay basin; Notre Dame Bay basin; Bay d' Espoir basin; Trinity Bay basin; and Conception Bay basin.

PHYSICAL DIVISIONS

Newfoundland may be divided into upland and lowland. Examine the map of Fig. 13.1. The upland over large areas has remarkably little relief, and generally breaks off in steep slopes to the lowland. Most lowland areas are on weak rocks, and a number of the steep slopes between upland and lowland are known to be fault-line scarps; others are thought to be. An article of Twenhofel and MacClintock (1940) discusses the physiography of Newfoundland and is the basis for the following review.

The highest part of the upland is the Long Range topographic feature along the west margin of Newfoundland. It has been referred to as a

plateau because of a fairly flat top. Actually only remnants of a high, flat surface exist, and these are about 2000 feet above sea level. High valleys of late mature aspect range in elevation from 1300 to 1700 feet and are correlated with the highest surface in the central plateau at 1400 to 1600 feet. This same surface declines to about 1000 feet in the Baie d'Espoir region, and 700 to 800 feet at St. Johns. A third surface in the western Long Range is at 500 to 1000 feet above sea level. In the central plateau this surface is believed to mark the mature upland of 500 to 1000 feet at Grand Lake—White Bay basin, and the 200-foot level at Notre Dame Bay and the lower Exploits basin, and the 350- to 400-foot level at St. Johns. The three surfaces, or peneplains, are regarded as sloping to the east and representing corresponding tilt of the island in that direction. The peneplains were developed through fluvial erosion, not marine; and as in the southern and central Appalachians were eroded, it is believed, in Tertiary time. Perhaps the highest Long Range peneplain formed in the late Cretaceous.

The Anguille Mountains have an upland surface much like that of Long Range. The highest points are at about 1800 feet above sea level. The Serpentine Range includes the Lewis Hills and Blow-me-down Mountains south of the Bay of Islands, Arm Mountain on the north side of the bay, the St. Gregory highland on the north entrance of the bay, Table Mountain on the south side of Bonne Bay, and Lookout Hills on the south entrance of Bonne Bay. These several relief features are parts of a basic intrusive complex. Lewis Hills have a remarkably flat surface at about 2300 feet above sea level, a well-preserved, mature surface at 1300 to 1700 feet, and a surface shown by upland valleys at 700 to over 1100 feet.

STRATIGRAPHY

Introduction

The stratigraphy of western Newfoundland was first summarized by Schuchert and Dunbar (1934). The report also reviews the stratified units of the rest of the island in the light of information up to 1934. Several *Bulletins* and *Information Circulars* of the Newfoundland Geological Survey under the direction of A. K. Snelgrove contain additional information; and a few journal articles by Twenhofel (1947), Twenhofel and Shrock (1937), Dorf and Cooper (1943), Kindle and Whittington (1958), and others present new stratigraphic and paleontologic information.

The island has been divided into four sections in the chart of Fig. 13.2 for the purpose of listing representative sequences. A fifth section is added on the west for the coast of Labrador, and still a sixth for the Canadian Shield. The chart attempts to summarize not only the stratified sections, but also the tectonic history. It can be referred to later when the structure and tectonic history of the island are discussed. The sections from west to east may represent the major stratigraphic provinces, since they are taken across the strike of the linear structural elements. The Notre Dame Bay section in the north-central part of the island and the Fortune Bay section in the south-southeastern part of the island may be parts of a common central province, the details of which are not yet known.

Cambrian System

In western Newfoundland limestones, dolomites, siltstones, and shales predominate and build up a sequence 3000 to 3500 feet thick. Along the west coast for a distance of 800 miles, and especially at Cows Head (between Bonne Bay and St. John Bay, Fig. 12.1) a succession of limestone conglomerates interbedded in shales and limestone, about 1000 feet thick represent Middle Cambrian to Middle Ordovician time (Kindle and Whittington, 1958). The conglomerates consist of small, flat chips, angular to subangular boulders, and scattered large blocks up to 600 feet in length. The matrix is a mudstone. The fragments came from a source area where calcarenites, oolites, calcilutites, and dense fine-grained, varicolored porcellanous limestones, in places with shale partings, were accumulating. The boulders have fossils of the same age as the matrix. These observations lead Kindle and Whittington to conclude that the conglomerates are not thrust breccias but intraformational units in a flysch sequence. The source direction could not be determined.

The Burin Peninsula has Cambrian beds of carbonates, shale, and

sandstone plus manganiferous shales and limestones, and about 1000 feet of sandstone and shale in the Conception Bay area. In the Rencontre East area of Fortune Bay, a section of Lower Cambrian or Proterozoic rocks is composed of more than 6000 feet of conglomerate, sandstone, arkose, limestone, and shale. So far, no volcanic rocks have been recognized in the Cambrian in Newfoundland.

In the Conception Bay area, it is interesting to note the occurrence of a great volcanic series, the Avalon, that underlies the Cambrian. Within the Avalon volcanic series at least three Precambrian epochs of sedimentation and volcanism are recognized, and each was terminated by folding, uplift, and erosion. The last disturbance probably preceded the deposition of the Cambrian only a short time, and the whole of the Avalon peninsula probably sank thereafter and was covered by the Cambrian sediments.

The fossils of all Cambrian sections have European affinities.

Ordovician System

The Ordovician strata of western Newfoundland consist of a lower sequence of 6700 feet of sandstones, shales, limestones, and dolomites, and an upper sequence, some 5000 to 10,000 feet thick, of dark and variegated shales and sandstones with minor amounts of conglomerate, arkose, and limestone. Some lava flows, agglomerate, and ash beds have also been noted in the upper or Humber Arm series. These are the first evidence of volcanism in western Newfoundland, and they were probably extruded near the close of the Ordovician.

The two thick sequences are separated by a disturbance that involved considerable faulting and erosion. The lower is massive and more competent; the upper is generally thin-bedded and incompetent. It is much distorted in nearly all outcrops. The volcanics in the upper sequence probably preceded ultramafic serpentine intrusions that penetrate the beds extensively.

The Ordovician in the Notre Dame Bay and in Fortune Bay areas is replete with volcanics. The sequences are very thick and generally associated with clastics containing the impure varieties of sandstone—arkose and graywacke. Only at the base of the Ordovician section, in the Fortune

	CANADIAN SHIELD	COAST OF LABRADOR	WESTERN NEWFOUNDLAND	NOTRE DAME BAY AREA	FORTUNE BAY AND BURIN PENINSULA	AVALON PENINSULA
?			FOLDING, THRUSTING	FAULTING	FAULTING	FAULTING
PENN.			BARACHOIS SER. CONGL., SS., SH. COAL, 3,000'			EROSION
PENN.		SHARP UPLIFT		UPLIFT	UPLIFT	UPLIFT
MISS.	GENERAL EROSION	EROSION	CODROY SERIES CONGL, SS, SH, LS. GYPSUM ANGUILLE SER CONGL, SS, SH. 3,500'	SPRINGDALE GROUP RED CLASTICS, VOLCANICS		EROSION
DEVONIAN		SHARP UPLIFT	DISTURBANCE INTRUSIONS	SHARP UPLIFT	BATHOLITHIC INTRUSIONS	UPLIFT
DEVONIAN		EROSION	CLAM BANK SERIES COARSE CONGL, SS., 1,700'	? ? ?	GREAT BAY DE L'EAU CONGL 3,000'	?
SILURIAN		SHARP UPLIFT	FOLDING, EROSION ?	FOLDING, FAULTING INTRUSIONS	FOLDING, EROSION?	SHARP UPLIFT
SILURIAN		EROSION	?	SILURIAN IN WHITE BAY, CLASTIC, 2,800' SILURIAN IN NOTRE DAME BAY, CLASTIC 2,000'	RENCONTRE FM QTZ, GRAYWACKE, VOLCANICS, 3,500'	?
SILURIAN			ULTRABASIC INTRUSIONS	DISTURBANCE ?	DISTURBANCE ?	?
ORDOVICIAN	GENERAL UPLIFT	? UPLIFT ?	GREEN SS. COW HEAD GROUP	HUMBER ARM SERIES WITH RED CLIFF VOL. GROSS POND VOLCANICS SNOOKS ARM VOLCAN TABLE HEAD SERIES SH. LATE ST GEORGE SER SLATES, VOLCANICS	MOORING COVE VOLCANICS, 1,500' ANDERSON COVE SLATES 1,500 BELL BAY VOLCANICS 13,000'± POOLS COVE CONGL 5,000'	? WABANA SH. HEMATITE 3,000' BELL ISLE 6.000'
ORDOVICIAN					DISTURBANCE	
ORDOVICIAN					BAY D'EST LS. 2,000'	CLARENVILLE SH. SS
CAMBRIAN		?		?	DISTURBANCE AND EROSION	JOHANNIAN 500'
CAMBRIAN		?		?	KELLIGREWS SH LONG POND SH CHAMBERLAINS BR SH	MANUELS SH. 300'
CAMBRIAN					EROSION	
CAMBRIAN					HANFORDIAN SH., LS.	HANFORDIAN SH., LS
CAMBRIAN					EROSION	EROSION
CAMBRIAN		LABRADOR SER. SS., LS., 470'	LABRADOR SERIES SS., LS., QTZ., 2,600'		BRIGUS CLASTICS AND LS.	ETCHEMINIAN SERIES CONGL., SS., SH., LS., 200'
PROTEROZOIC		UPLIFT, EROSION	UPLIFT, EROSION		FOLDING, INTRUSION	FOLDING, INTRUSIONS
PROTEROZOIC	EROSION	EROSION	EROSION	EROSION	BAY D'ESPOIR SERIES, 15,000' GRAYWACKE, BASAL VOLCANICS HARBOUR MAIN VOLCANICS OF BURIN P	AVALON VOLCANIC SERIES 15,000' TWO OROGENIES WITHIN SERIES

Fig. 13.2. Representative sections and crustal disturbances of Newfoundland. Compiled from various reports mentioned in the text and with the aid of Daniel A. Bradley, University of Michigan. The age of the folding, faulting and intrusions of the Notre Dame Bay area as indicated between the Silurian and Mississippian beds is doubtful; they may be Acadian rather than Caledonian.

Bay area, is a nonvolcanic series present. There about 2000 feet of limestone occur.

In Conception Bay on the east the volcanics are absent, or if deposited, have been eroded away. The Belle Isle and Wabana formations, the latter carrying sedimentary iron-ore beds, are chiefly sandstones and shales, about 9000 feet thick.

The thick Ordovician sections in central Newfoundland with their abundant volcanics resemble the Ordovician Ammonoosuc volcanics of New Hampshire more than any strata of similar age in the Maritime Provinces.

Silurian System

Strata of Silurian age are not known in either the western or eastern divisions of Newfoundland, but in the central belt various clastics are fairly voluminous. In the White Bay and Notre Dame Bay areas over 2000 feet of Silurian sandstones and shales have been noted. In the Fortune Bay area, the Rencontre formation consists of quartzite, graywacke, and volcanics, about 3500 feet thick. Other sequences in the central division may prove to be Silurian.

Devonian System

The Clam Bank series along the western shore of St. George peninsula is a coarse, red conglomerate, with intercalated masses of soft, coarse brown sandstone and shaly sandstone of early Devonian age. The well-rounded and polished pebbles in the conglomerates are of many kinds and range up to 4 inches in diameter. The beds resemble the Triassic sediments of the Connecticut Valley. In places they appear nearly flat, but in others they are on end. They indicate a sharp uplift immediately preceding and collateral with their deposition, and their deformation indicates a following orogeny.

In the Fortune Bay area, the Great Bay de l'Eau conglomerate is 3000 feet thick, and is also believed to be early Devonian.

Early Devonian plant impressions were discovered in the La Poile Bay area of southeastern Newfoundland east of Long Range in 1940 (Dorf and Cooper, 1943) in the Bay du Nord series which, because of its metamorphosed character, had previously been thought of as Precambrian. The fossils occur in a grayish-black slate which is associated with graywacke and conglomerate. Much of the central plateau is metamorphic and igneous rock, and a belt of schistose character flanks Long Range on the east. The early Devonian fossils occurring in rocks of this terrane open up the possibility that much of the stratified altered rock, previously called Precambrian, is Paleozoic; and that the numerous and large cross-cutting plutons are Acadian in age. Recognizing the well-established Acadian orogenic history in the Maritime Provinces and in New England, which includes much metamorphism and plutonic activity, a number of modern investigators are classifying the stratified, altered, lithologic units as Paleozoic rather than Precambrian. It seems probable that much of central Newfoundland will prove to be underlain by Paleozoic rocks. The recent *Geologic Map of North America* shows most of it as Ordovician strata and Devonian intrusives. Undoubtedly more Paleozoic systems will be recognized in this complex in future investigations.

Mississippian System

Mississippian rocks are present abundantly in the St. George Bay area and in the White Bay—Grand Lake lowland. They are also known at Cape Rouge and Groais Island, and in part of the Notre Dame Bay area. The chart of Fig. 13.3 correlates the Mississippian formations of these areas. They are chiefly clastics. The St. George Bay series contains in addition some evaporites, and the Notre Dame Bay area, some volcanics. Up to 3500 feet of beds have been noted in these sections.

ST. GEORGES BAY AREA		DEER LAKE	WHITE BAY	CAPE ROUGE - GROAIS ISLAND	RED INDIAN LAKE	NOTRE DAME BAY
CODROY SERIES	WOODY POINT SS.	UPPER GRAY AND LOWER RED SHALES			SHALE, CONGL. AND LIMESTONE	
	WOODY COVE SH,LS					
	BLACK POINT LS.					
	CODROY SH., GYP.			CAPE ROUGE SH., SS., SILTSTONE		
ANGUILLE SERIES	SNAKE BRIGHT SH.		SPEAR POINT SS.,SH.	?		SPRINGDALE GRP. RED CLASTICS, VOLCANICS
	CAPE ANGUILLE SS.			PILIER CONGL.,SS.		

Fig. 13.3. Mississippian formations of Newfoundland, after Betz, 1948. All are regarded as Lower Mississippian.

Pennsylvanian System

A body of coarse clastics, the Barachois series, rests on the Mississippian Codroy series in the St. George Bay area. It consists of 5000 or more feet of coarse conglomerate, sandstone, arkose, and shale, with some thin coal beds, presumably all continental, and indicates a new sharp uplift nearby. No other Pennsylvanian strata are known in Newfoundland.

INTRUSIONS

Serpentine Belts

Two belts of ultrabasic plutons occur in Newfoundland. They are known as the eastern and the western serpentine belts. Not only serpentine but also chromite are common associates of the basic intrusions (Snelgrove, 1934). The principal rocks are peridotite, pyroxenite, and gabbro. See map, Fig. 13.4.

The eastern serpentine belt extends from Carmanville to the headwaters of the Gander River. Serpentine masses are exposed intermittently over 120 miles in a general northeast-southwest direction. According to Snelgrove:

> This part of the island is relatively low-lying and is characterized by undulating topography. The ultrabasic rocks of this belt, in contrast with those on the west coast, are only partly exhumed by erosion and consequently lack any striking topographic expression. The serpentines form low, bare ridges, with few prominent peaks or knolls.
>
> At the north tip, it has an outcrop width of one-half mile, and dips westward. Highly serpentinized dunite is confined to a band varying from one hundred to five hundred feet in width, flanked by pyroxenite. The serpentine band forms small prominences. The country rocks beneath the intrusives are chloritized volcanics, locally fragmental, underlain by micaceous black slate and quartzite. Above the ultrabasic rocks are black slate, gray quartzitic sandstones, and conglomerate. These sedimentary and volcanic rocks are probably of early Paleozoic age; they appear to have been intruded conformably by the plutonic rocks.
>
> The section of the belt exposed near the headwaters of the Gander River, central Newfoundland, consists of serpentinized dunite with lenticular segregations of medium-grained to pegmatitic pyroxenite. Its width was not determined. Structurally, the intrusion appears to be nearly vertical; it is invaded by a granite batholith lying to the south and east.
>
> The Western Serpentine Belt consists of a series of four main intrusions, which seem to have been injected concordantly at different horizons into a

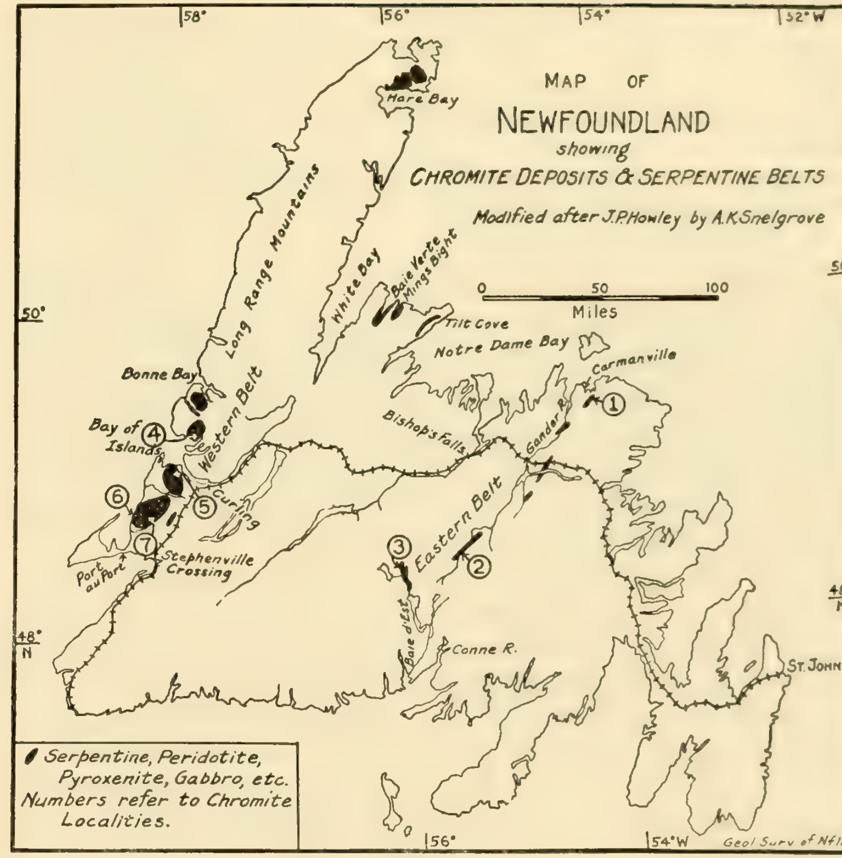


Fig. 13.4. Ultramafic plutons of Newfoundland. Reproduced from Snelgrove, 1938.

> folded sedimentary and volcanic series (Humber Arm series), probably of upper Ordovician age, which underlies this part of the lowland of the west coast of Newfoundland.
>
> South of Bay of Islands, the eastern section of this belt, as exposed in Blow-me-down Mountain, is a pseudo-stratified complex and is composed of a wide zone of various types of peridotites at the base, succeeded by more siliceous rocks toward the top. Both the intrusives and the country rocks of sandstones, slates, argillites, and lavas have a general westward dip near Blow-me-down Mountain. In the section south of Bay of Islands, a lopolithic structure is indicated. Five miles to the east of the southernmost intrusive of the western belt is a satellitic serpentine mass containing an asbestos prospect. The structural

relations of the mass are unknown. A smaller satellite some 1,000 feet thick and well-differentiated occurs in Lark Mountain, south of the mouth of Bay of Islands.

North of Bay of Islands, also, the basal portion of the ultrabasic rocks composing the serpentine belt is composed of a wide zone of peridotites which dip westward.

Since no igneous rocks are known to cut the Carboniferous of western Newfoundland, the intrusives are referred to either the Taconic (late Ordovician) or the Acadian (late Devonian) orogeny.

The western serpentine belt extends adjacent to the west coast from Port au Port Bay to Bonne Bay and forms the flat-topped Serpentine Range, previously mentioned, with summit elevations around 2000 feet.

Other areas of serpentine not included in the eastern and western belts are on the east side of the northern peninsula at Hare Bay, and at Baie Verte and Ming's Bight. At Hare Bay considerable thicknesses of peridotites have an eastward dip and, with the enclosing sediments, form the eastern limb of the northern peninsula anticline.

At Baie Verte, the formation of that name, which consists of greenstone and greenstone schist with minor amounts of graywacke, tuff, agglomerate, lava, slate, ferruginous chert, sandstone, and marble, has been intruded by large, dominantly concordant bodies of ultramafic rock and gabbro (Watson, 1943). Much of the ultramafic rock has undergone intense serpentinization and steatitization. The gabbro has suffered saussuritization, uralitization, silicification, carbonatization, and alteration to zoisite-quartz and zoisite-prehnite rock. Granite, quartz-porphyry, and quartz-diorite intrusions occur in the Baie Verte formation. Adjacent to the latter, the greenstone and gabbro have been metamorphosed to the amphibolite facies. Small sills and dikes of mafic gabbro, porphyrite, diorite, and kersantite were observed in the area.

The above areas of ultramafic rocks are shown on the map of Fig. 13.4. These occurrences in Newfoundland are considered part of a major serpentine belt from Georgia through the crystalline piedmont belt to the Hudson Valley and through the Taconic system to the St. Lawrence and the Gaspé Peninsula. They have been compiled by Hess, and his map is reproduced in Fig. 8.29. Hess has developed the theory that serpentine plutons occur in linear arrangement and mark the heart of the belts of great compressional deformation, especially of the volcanic arc type. If the linear belt of ultramafic plutons be interpreted in this way, we have to deal with additional evidence of a great orogenic belt, and can point to its core of greatest deformation.

In the St. Lawrence–Gaspé belt, most of the serpentinized plutons are Taconic in age, but some may be Devonian. About the same can be said of their age in Newfoundland. Their age is not known in the crystalline piedmont, but it is inviting to think of the entire serpentine belt as one of the manifestations of the great Taconic orogeny.

Granitic Plutons

Many large discordant granitic to dioritic plutons, some of batholithic proportions, occur in the central part of Newfoundland between the Precambrian of Long Range and the Precambrian of Avalon peninsula. Some lie within the Precambrian areas also. For the most part they have not yet been mapped and differentiated. They are now regarded as probably Acadian in age, since one has been found intruding the early Devonian beds of the La Poile Bay area and another one cuts the Devonian beds of the Fortune Bay area. Some may be late Silurian (Caledonian); most are known to cut the Ordovician strata, and pebbles of the granite are found in a Mississippian conglomerate.

Instructive examples are the Bay du Nord granodiorite and Ackley granite batholiths of the Fortune Bay and Burin peninsula region. See map, Fig. 13.5. According to White (1940):

The (Ackley) batholith intrudes the northwest limb of a large syncline, the major structure of the Fortune Bay synclinorium. The invaded rocks are largely the Ordovician (?) Belle Bay volcanics, and to a lesser extent, tuffaceous slates conformably overlying the volcanics, and Cambrian quartzites. The mapped extent of the batholith is over 160 square miles, but this is probably less than half of the total. The long axis of the intrusion is oriented approximately northeast, parallel to the dominant regional structural trends. The dip of the contact, where it could be determined, is 25° to 45° outward from the batholith.

The topography of the batholith is of low relief, with elevations averaging about 750 feet, in contrast to the higher elevations and considerably greater local relief of the volcanics to the south.

The intrusion consists mainly of granite ("white granite") and alaskite ("red-granite"), with the latter the more abundant, in the southern part of the batho-

lith. These two phases are generally gradational, but sharp contacts and local cross-cutting relationships have been observed.

Basic and intermediate rocks are completely absent, although early phases of the differentiation series may be represented by the nearby Bay du Nord batholith.

The Bay du Nord and Ackley batholiths are in turn cut by the Belleorum granite, which is known to intrude the Great Bay de l'Eau conglomerate of Devonian age (D. A. Bradley, personal communication). The three plutons are regarded by Bradley as closely related genetically.

Composite batholiths have been noted in the St. Lawrence area of the Burin peninsula where the Lawn (?) metagabbro, possibly of Taconic age, is succeeded by the St. Lawrence granite of Acadian age (Van Alstine, 1948); in the Trinity Bay area where the Powder Horn diorite is intruded by the Northern Bight granite (Hayes and Rose, 1948); and in the Notre Dame Bay area where a pink granite batholith with satellites in the Hodges Hills vicinity intrudes a gray hornblende diorite. The latter diorite has gabbro facies and exhibits all the characters of xenolithic assimilation (John J. Hayes, personal communication).

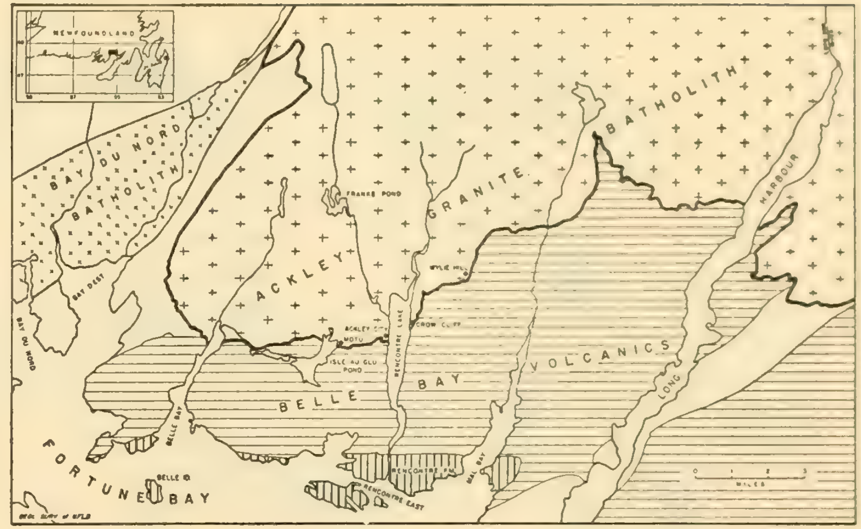


Fig. 13.5. Geologic map of Recontre Bay area. Reproduced from White, 1940.

MAJOR STRUCTURAL DIVISIONS AND THEIR CHARACTERISTICS

Tectonic Map

The tectonic map of Fig. 13.6 is an attempt to classify the major structural divisions of Newfoundland, and to show some of the important fold axes and faults of the large island. It is based chiefly on Snelgrove's *Geologic Map of Newfoundland* (1938) and on additions that he has made on a copy loaned to the writer. The faults and folds of the Notre Dame Bay area were taken from a work map of John J. Hayes.

Considerable field work has been done that is not yet in print; much of the central plateau has never been seen by geologists; and areas of crystalline rock are now being considered more as Acadian orogenic complex rather than Precambrian. These factors lead to an almost hopeless task of bringing the geologic map up to date and making it tolerably correct, even if generalized. As a substitute, a generalized tectonic map was constructed (Fig. 13.6) that divides Newfoundland into four major geologic zones, each with distinguishing characteristics. In addition, the Carboniferous basins, basic plutons, principal fold axes and faults, and Cambrian outcrops, as far as known, are shown. Each zone will be described separately.

Principal Structural Directions

Overall, the fold axes, the faults, and the foliation take a north-northeasterly direction; but upon closer observation, some structures trend more easterly, especially in the Notre Dame Bay area. The stratigraphic and structural composition is much like that of the Maritime Provinces and New England, and undoubtedly Newfoundland is part of the great Appalachian Mountain systems.

Relation to Physiographic Provinces

The Long Range highland of the physiographic map, Fig. 13.1, is coincident with the crystalline Precambrian (?) rocks of zone 1 of the tectonic map, Fig. 13.6; the serpentine plutons are generally strong relief

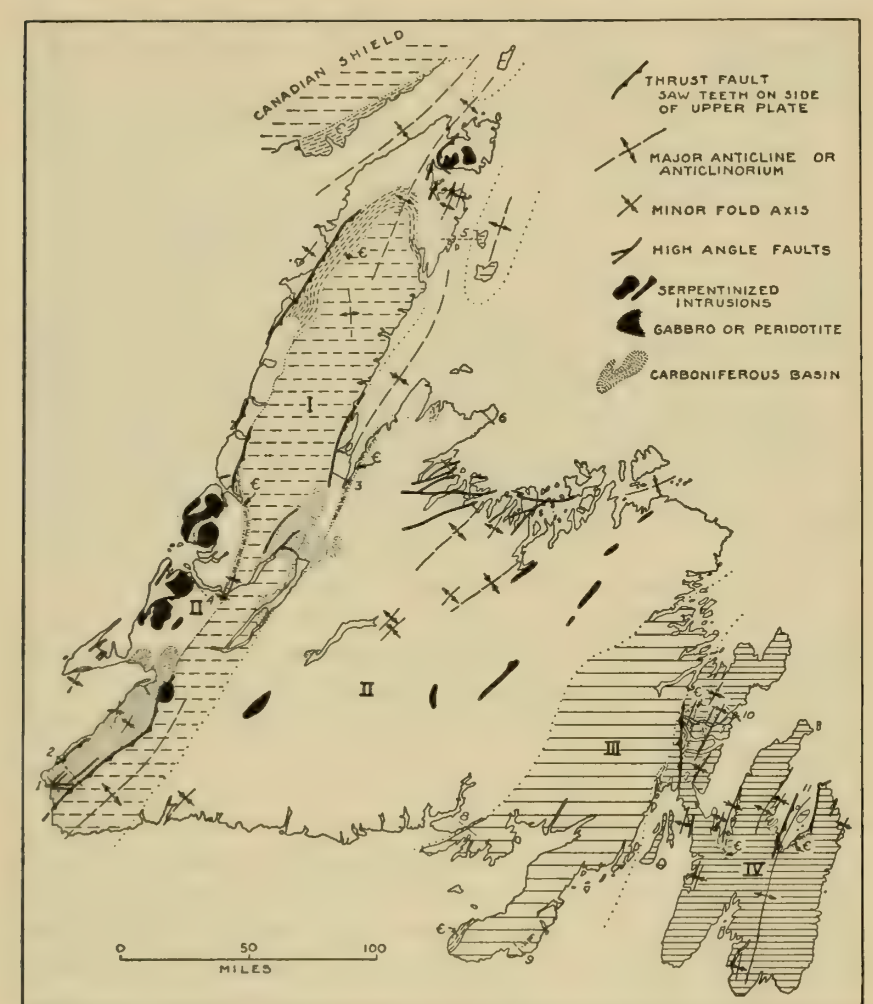


Fig. 13.6. Tectonic map of Newfoundland, taken mostly from Snelgrove's *Geologic Map of Newfoundland*, Newfoundland Geological Survey. Interpretations assisted by J. J. Hayes, D. Bradley, and Joe Kerr. Zone I consists of schists, gneisses, and intrusives, believed to be chiefly Precambrian, which in part may be metamorphosed volcanics. It was actively deformed during Taconic and Acadian orogenies. Zone II is the Paleozoic orogenic belt of Ordovician, Silurian, and Devonian metasediments, metavolcanics, and batholiths. It may contain both older and younger rocks, but in exposure they are of minor importance. Zone III is a Paleozoic orogenic belt, but in addition to the rocks of zone II it contains major Precambrian linear elements. Zone IV consists principally of Precambrian sediments and volcanics with small infolded or faulted basins of Cambrian and Ordovician strata. The zone is generally much less deformed than the others. Carboniferous basins are stippled and postdate the major orogeny, but were affected by Appalachian faulting. Black areas with smooth borders are serpentinized intrusions, and black areas with hachured edges are gabbros and peridotites. Numbers 1 to 11 are lines of cross sections.

features; and the Carboniferous areas are for the most part lowlands; but the uplands and lowlands east of these do not clearly indicate individualized geologic provinces, as far as known.

Characteristics of Tectonic Zones

Zone One. Zone one is the Long Range highland, and it consists chiefly of schists and gneisses similar to those of the nearby Canadian Shield of Labrador. At the south and between La Poile Bay and Cape Ray, however, part of the rocks may be metamorphosed Paleozoic. George Phair has mapped the coast from La Poile Bay westward, according to Joe Kerr (personal communication), and finds at the bay a fossiliferous Lower Devonian formation with the argillaceous members slaty and sharply folded. As the upturned succession is traversed westward, it becomes phyllitic and finally schistose. No contacts could be found between the Devonian slates and the phyllites, and the schists, previously called Precambrian. Phair visualizes the southern end of the Long Range as an anticlinorium of isoclinal folds, pitching north-northeastward, and with increasing metamorphism toward the core; perhaps Precambrian rock is exposed in the core, but contact relations are not evident to prove it.

At the north end of the range and along its flanks at intervals—Bay of Islands area on the west and White Bay on the east—Cambrian beds rest on the schists and gneisses, and hence demonstrate the Precambrian age there of the foliate rocks.

Zone Two. Zone two east of Long Range appears to be basically the Acadian orogenic complex. It is made up principally of the great Ordovician and Silurian volcanic sequences and numerous great batholiths, presumably of Caledonian or Acadian age. The stratified sequences are much folded and generally subject to low-grade metamorphism. Some Precambrian rocks may exist, but this possibility seems less as work

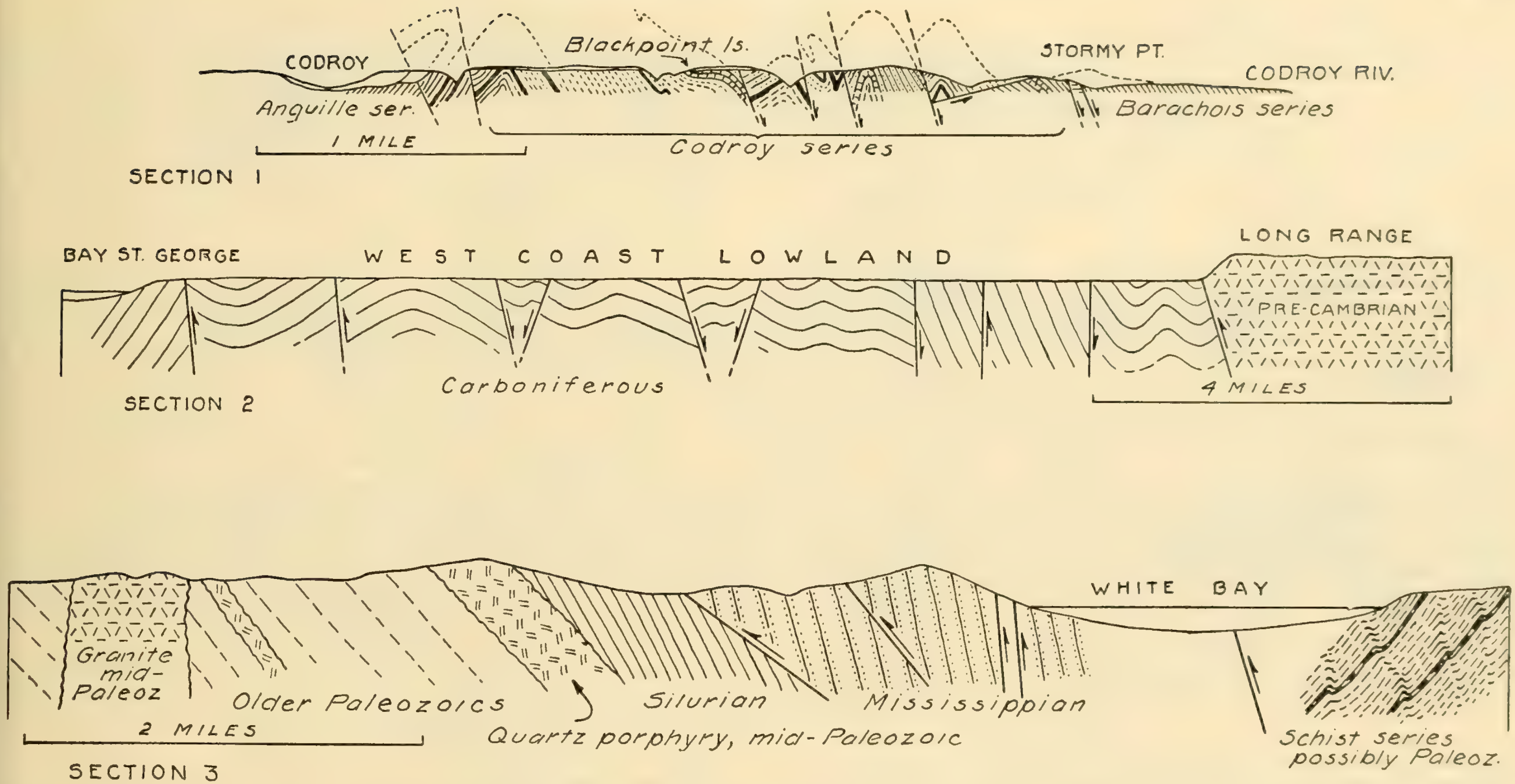


Fig. 13.7. Representative cross sections of Newfoundland. Section 1 after Hayes and Johnson, 1938; section 2 after Betz, 1943; section 3 after Betz, 1948.

progresses. Much of the region is unknown. Several serpentinized ultramafic intrusions occur in a line southwest of Carmanville.

Zone two west of the Long Range Mountains consists of folded and faulted Cambrian, Ordovician, and Devonian strata, with the Ordovician thickest but with volcanic rocks present in only one formation. It represents the front of the Taconic and Acadian systems. It contains the major Carboniferous basin and the principal belt of serpentine intrusions.

The Long Range has been elevated in a steep reverse fault against the Carboniferous basin. See section 2, Fig. 13.7. Section 1 shows the faulted and folded nature of the Carboniferous rocks themselves. They are generally far less folded, however, than the underlying Ordovician. Folded Carboniferous is also shown in section 4B resting unconformably on the

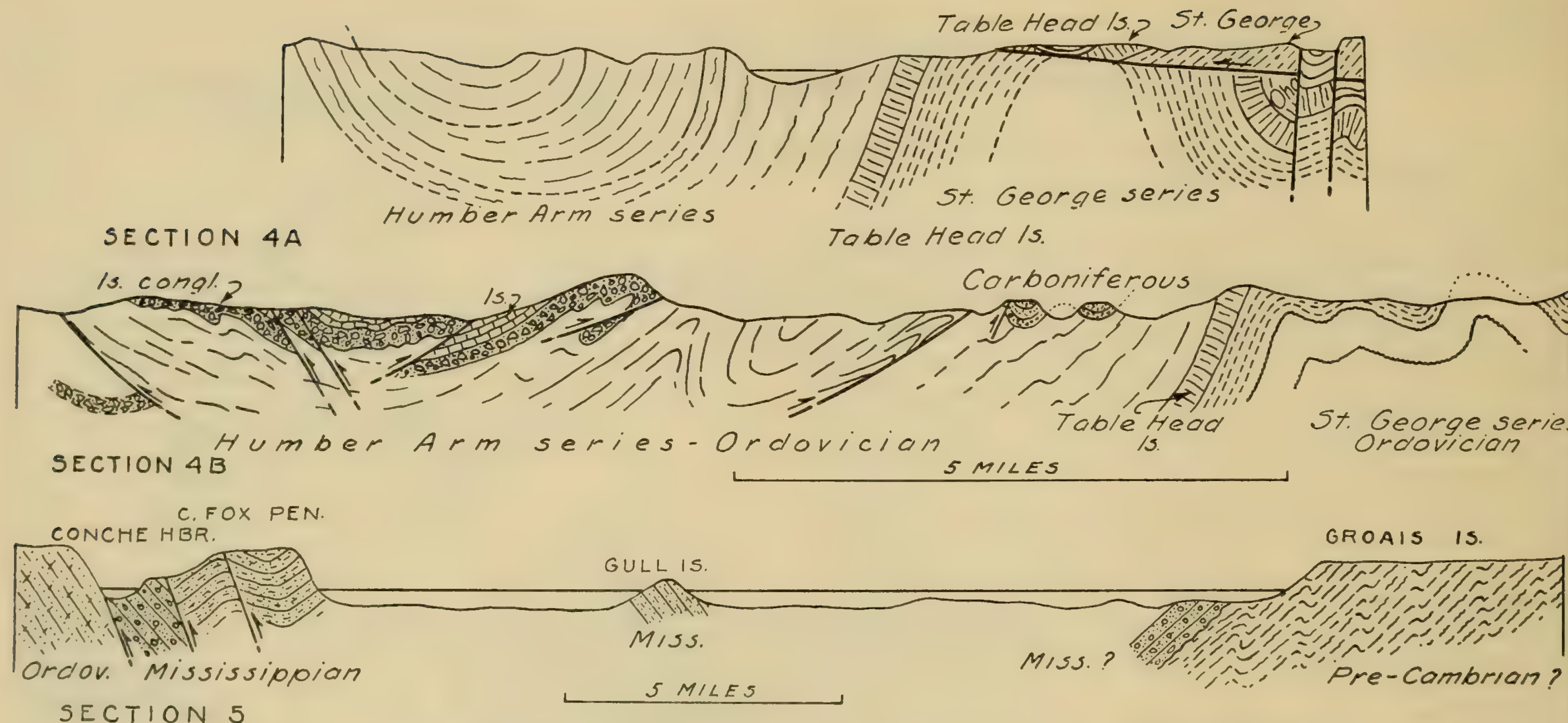


Fig. 13.8. Representative cross sections of Newfoundland. Section 4A and 4B after Walthier, 1949; section 5 after H. Johnson, 1941.

Humber Arm series of the Ordovician.

The upfaulting of Long Range on the west started in early Mississippian time and resulted in the deposition on the downfaulted block of the coarse Anguille series. Movement continued during the deposition of the entire Mississippian and Pennsylvanian sequence, or at least recurred after the Mississippian sediments were deposited, because the Precambrian is now in fault contact with the Mississippian. Faulting recurred after the Pennsylvanian Barachois beds were deposited.

The structure along the east flank of the Long Range uplift is illustrated in sections 3 and 5, Figs. 13.7 and 13.8. High-angle thrust faulting seems the dominant structure, but probably a large syncline or synclinorium exists between the mainland and Groais Island. Groais and Bell islands are presumably Precambrian schists and gneisses, and hence are believed to mark an anticlinal fold.

Representative of the folding and faulting in the Notre Dame Bay area are sections 6 and 7 of Fig. 13.9. Through the islands and headlands of Notre Dame Bay area, a system of faults with an east-west bearing occurs. Those shown on the tectonic map were taken from a compilation

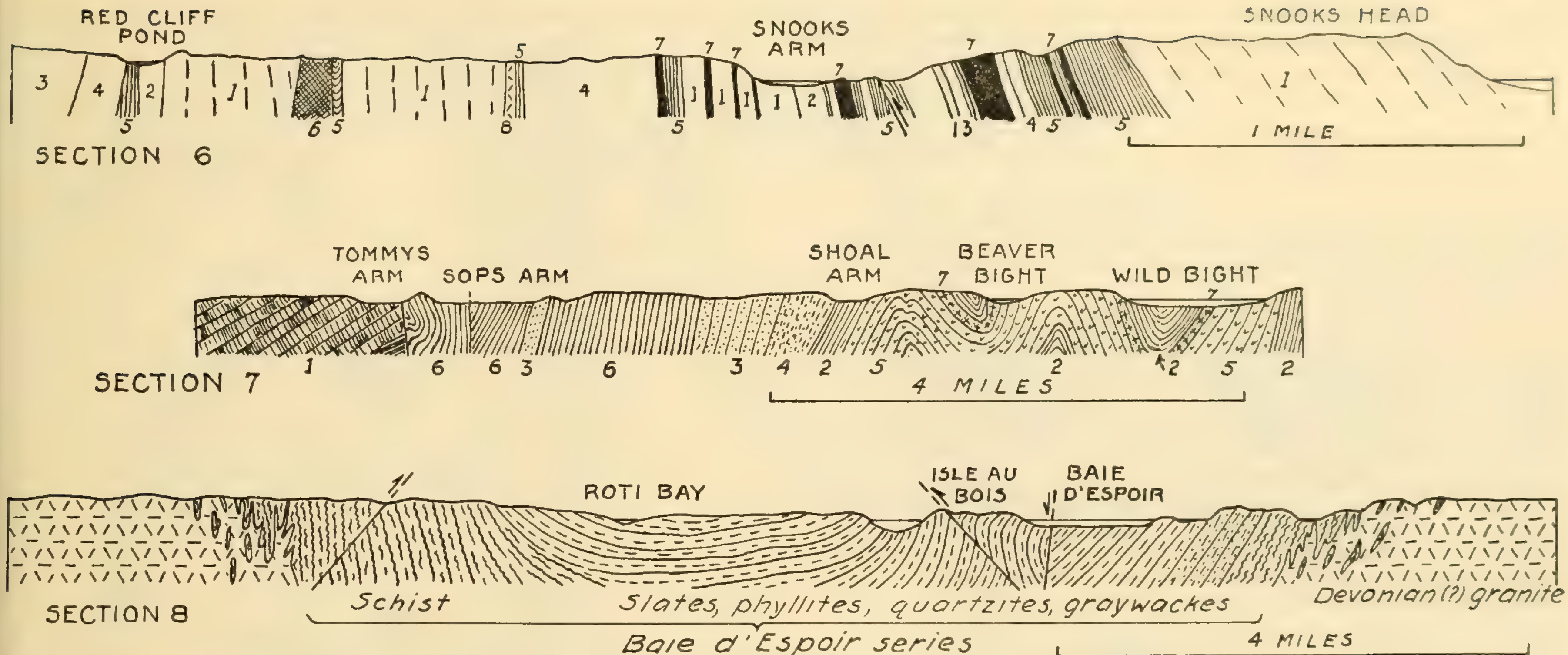


Fig. 13.9. Representative cross sections of Newfoundland. Section 6 after Snelgrove, 1931; 1 to 5 make up the Snooks Arm series of Ordovician age. 1, andesite pillow lava; 2, andesite; 3, rhyolite; 4, pyroclastics; 5, slates, argillite, sandstone, chert. Nos. 6 to 8 are post-Ordovician. 6, gabbro; 7, diabase and basalt; 8, Burtons Pond granite porphyry. Section 7 after Espenshade, 1937. 1, pillow basalts; 5, andesites; 4, shales and sandstone; 3, coarse, massive sandstone; 6, argillaceous graywacke and chert; 2, shales, tuffs and cherts; 7, gabbro. All units are probably Ordovician. Section 8 is after Jewell, 1939.

by J. J. Hayes. Some of the northeast are probably horizontal shears, and the main east-west faults are high-angle ones with movement in the vertical direction. The fold axes trend acute to the major faults, and to put them in the same mechanical frame as the folds seems impossible. The folds appear to the writer to be Acadian, and the faults more likely to be associated with the faulting of the Carboniferous basins and later than with the Acadian folding.

Zone Three. Zone three is much like zone two but includes several Precambrian linear masses. These may be upfaulted blocks or cores of anticlinoria. Cross sections 8, Fig. 13.9 and 9A and 9B, Fig. 13.10, are representative of the structure. They show especially the transgressive granitoid intrusions. The Precambrian rocks that appear in zone three are sediments and volcanics, and are considered later than the schists and gneisses of Long Range.

Zone Four. Zone four is predominantly a late Precambrian sedimentary and volcanic series, with infolded or downfaulted Cambrian and Ordovician sediments in several places. On Belle Isle of Conception Bay, Ordovician sediments occur which carry iron ore. See map and sections,

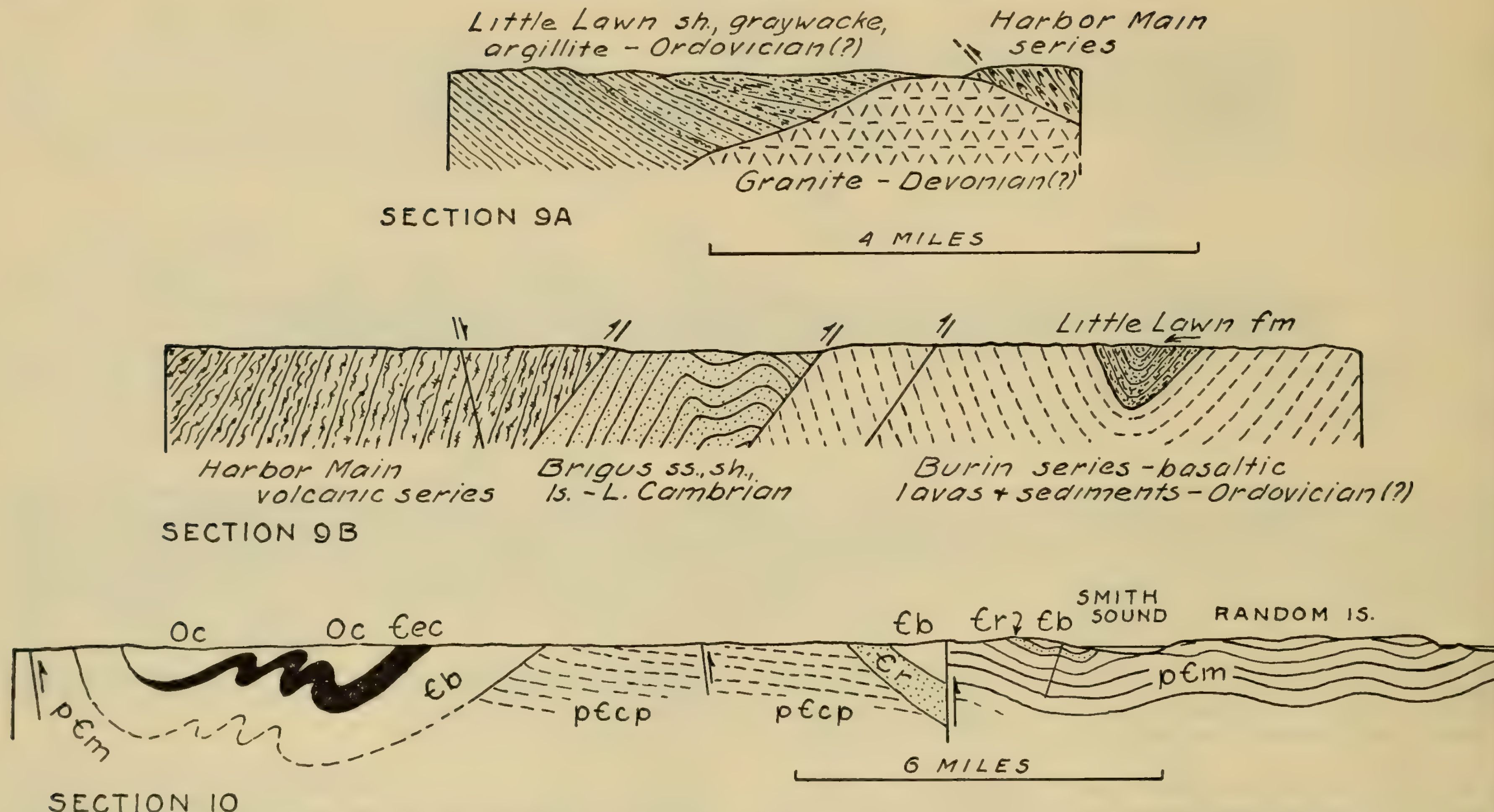


Fig. 13.10. Representative sections of Newfoundland. Sections 9A and 9B after Van Alstine, 1947; section 10 after Hayes and Rose, 1948; p€m, Musgravetown granite; p€cp, Connecting Point granite; €r, Randon quartzite; €b, Brigus conglomerate, quartzite, shale; €ec, Elliott Cove shale; Oc, Clarenville shale, sandstone.

(locality 11) Fig. 13.11. The Cambrian and Ordovician sediments have largely escaped metamorphism. Along the west side of Trinity Bay (section 10) the Cambrian and Ordovician sediments are rather tightly folded, whereas to the east in Avalon peninsula, the Paleozoic beds are less folded and chiefly faulted. The impression is conveyed that zone two east of the Long Range Mountains suffered the most intense deformation, and that zones one and three, although deformed and intruded extensively, are marginal; and that the eastern part of zone four escaped the

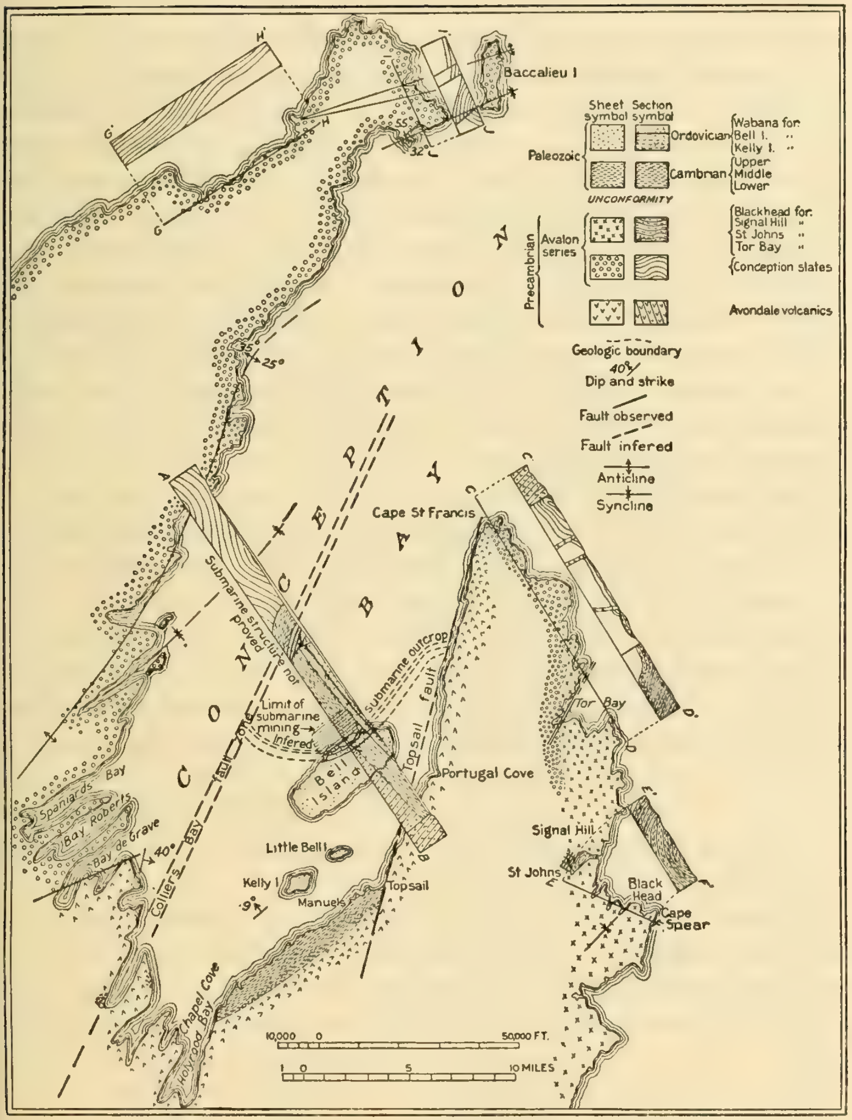


Fig. 13.11. Map and sections of Conception Bay and the Wabana iron ore deposits. Reproduced from Hayes, 1931.

sharp folding and metamorphism common to zones one, two, and three but was faulted and elevated in post-Ordovician time, probably in the Appalachian orogeny.

TECTONIC HISTORY

Early Cambrian Phase

By reference again to the chart of Fig. 13.2, the numerous disturbances and orogenies that characterized the Appalachian systems in Newfoundland can be reviewed. Nine orogenic phases are fairly clear. When correlations are more precise, this number may be increased.

It is evident from the angular unconformity at the base of the Cambrian and the coarse, basal clastics that an orogeny immediately preceded or accompanied the early Cambrian sedimentation. This is noted in the west along the coast of Labrador and the western lowland of Newfoundland, and in the east from the Bay d'Espoir to the Avalon peninsula. In the east the orogenic phase is the last of three or more that accompanied the deposition of a great Precambrian volcanic series. It is not yet possible to define the distribution of land and sea in the orogenic belt in Cambrian time. For that matter, the same can be said of the belt in all pre-Carboniferous time. Volcanic activity was pronounced in eastern Newfoundland in the Proterozoic but abated everywhere, it seems, during the Cambrian period.

After the early Cambrian Brigus and Etcheminian clastics had been deposited in the Fortune Bay and Trinity Bay areas respectively, a slight disturbance occurred which resulted in uplift and erosion before the next Lower Cambrian Hanfordian beds were deposited.

Late Cambrian Phase

In the Burin peninsula, a disturbance occurred in late Cambrian time in which the Middle Cambrian beds were tilted and somewhat eroded before covered with the Ordovician strata (Van Alstine, 1948). Deposition of carbonates occurred apparently undisturbed on the east and on the west while the uplift was taking place.

Early Orodovician Phase

With the beginning of Ordovician time, western Newfoundland started to sink more rapidly and became the site of deposition of a thick clastic series, and later of considerable limestone and dolomite. The central area around Fortune Bay received much limestone, at least in places. East Newfoundland also sank considerably and received over 6000 feet of fine clastics and carbonates. It seems necessary to picture the western Newfoundland Lower Ordovician clastics coming from the Canadian Shield where a rather sharp uplift set in (see Plates 2 and 3), but the source of the shales in eastern Newfoundland is not clear.

After early Ordovician time, the whole central part of Newfoundland became a site of profound volcanic activity, much of it submarine, with the passive emission of flows; but there was also abundant pyroclastic activity, probably both submarine and subaerial. The Ordovician must also have been a time of tumultuous crustal activity in the volcanic zone because various clastics, such as graywacke, conglomerate, sandstone, and shale, are commonly interbedded in the volcanics, or mixed with tuffaceous material, and they necessarily must have come from nearby uplifts. Chert and carbonate were also deposited, which with the above lithologies are the common associates of volcanic orogenic belts. In places upward of 20,000 feet of volcanics and sediments accumulated.

Andesites are the most common of eruptive rocks in the orogenic belts, but in the Belle Bay volcanic series of Fortune Bay, about 13,000 feet thick, most of the volcanics are rhyolite (D. A. Bradley, personal communication). This is indeed a great outpouring of rhyolite in an orogenic belt. Hobbs (1944) has found that andesites are the first eruptives in new orogenic belts in the southwest Pacific, but after a period of growth, other less basic forms appear, with rhyolite one of the late entrants. Since volcanic activity continued long after the Belle Bay rhyolites in central Newfoundland, it appears that new volcanic cycles followed the early Ordovician one.

Late Ordovician Phase (Taconic Orogeny)

The Taconic orogeny is generally held to have been pronounced in Newfoundland, not because of a great angular unconformity between Ordovician and Silurian rocks, but first, because the Ordovician sequences are more metamorphosed than the younger ones (Schuchert and Dunbar, 1934); second, because the Silurian has much conglomerate in it; and third, because the Taconic orogeny of the Gaspé and Maritime Provinces could not very well end abruptly without extension into Newfoundland. Silurian beds are relatively not very abundant in Newfoundland, and good exposures of their contact with the Ordovician sequences have so far escaped detection. Twenhofel and Shrock wrote in 1937 that so far as known there is no angular unconformity between the Ordovician and Silurian systems. However, White (1940 and Ph.D. thesis, Princeton, 1939) recognized evidence of the Taconic orogeny in the Rencontre East area of Fortune Bay, where the Long Harbour volcanics of Ordovician age were folded and extensively eroded, he believes, before the Silurian Rencontre series was deposited.

The contention that the Ordovician sequences are more metamorphosed than younger ones is correct only in so far as the "younger ones" are Carboniferous sequences or, perhaps in a few places, Devonian. Some of the granitic batholiths are now known to be Acadian, and most of the metamorphism may be incident to them, in which both Silurian and certain Devonian strata are altered as much as the Ordovician. Aside from the Rencontre East area, it is difficult to find tangible evidence of a sharp orogeny in Newfoundland at the close of the Ordovician. The Silurian series, with its volcanics and clastics, resembles the Ordovician of central Newfoundland, and it seems more logical to regard the central belt as one of continuing, but intermittent, orogenic and volcanic activity into and through the Silurian.

The ultrabasic intrusions of western Newfoundland are regarded as Late Ordovician mostly by relation to those of the Gaspé and Quebec Taconic belt (Snelgrove, 1934). Some of the ultrabasic plutons are known to intrude the Ordovician volcanic series and are therefore not older than the Taconic.

Late Silurian Phase (Caledonian Orogeny)

The Clam Bank conglomerate of western Newfoundland and the Great Bay de l'Eau conglomerate of Fortune Bay, both of early Devonian age,

indicate sharp uplift nearby, and the influx of much coarse clastic material. Since Devonian plant fossils have been found in schistose strata in the La Poile Bay area, it now seems probable that considerable of the metamorphic rocks of central Newfoundland, aside from the batholiths, will prove to be Devonian, and therefore a site of deposition during part of Devonian time, at least. The sources of the Lower Devonian conglomerates and sandstones must have been along the Labrador coast on the west and in an uplift through the Avalon peninsula on the east.

A Caledonian orogeny in the White Bay and Notre Dame Bay region has been suggested by Heyl (1937a) in view of the lithologic similarity of the Devonian and Mississippian there, in contrast to the Silurian and older rocks. Also, the amount of deformaton of the Devonian and Carboniferous is less than that of the older beds. Schuchert and Dunbar (1934) note that the Devonian sediments in the St. George Bay area are not strongly deformed, except along Appalachian phase faults; they are apparently no more disturbed than the Mississippian strata, and much less disturbed than the Ordovician Humber Arm series.

If an orogeny occurred in the White Bay and Notre Dame Bay region, it is not unlikely that intrusive activity accompanied the deformation. Some of the plutons of that region may, therefore, be Caledonian. They may also have come in during the Devonian or at its close (Acadian). Composite relations undoubtedly exist (Hayes, personal communication).

Late Devonian Phase (Acadian Orogeny)

Like the Taconic orogeny the Acadian is also illusive. Mississippian clastics in themselves indicate sharp uplift nearby, and are generally believed to rest in angular relation on much deformed Ordovician strata in western Newfoundland and in the White Bay and Notre Dame Bay area, although the contact is seen in only a few places. The Mississippian strata have suffered little metamorphism, however, and this sets them off strikingly from the older deformed and altered rocks. Nowhere in Newfoundland has an angular unconformity yet been recorded between the Mississippian and Devonian systems. Nevertheless, all workers in Newfoundland are aware of profound folding, batholithic intrusions, volcanism, and metamorphism that occurred sometime between the Ordovician and Mississippian; and since in two places the batholiths are found intruding the Lower Devonian series, it seems probable that many plutons, similar in composition, are of the same age. The Acadian orogeny, proceeding through the late Devonian and into early Mississippian in the Maritime Provinces and New England, was one of superior and widespread proportions, and it is highly unlikely that Newfoundland, with its similar geosynclinal assemblages and lying in the projection of the great belt of orogeny, could have escaped it.

Mississippian Phase

The desposition of the Anguille conglomerates in the St. George Bay area attended the upfaulting of the Long Range mass, and the same activity is probably indicated by the Pilier conglomerate at Groais Island. The Springdale clastics in the Notre Dame Bay area, if correctly dated, indicate orogeny nearby.

Early Pennsylvanian Phase

The coarse and thick Barachois series of the St. George Bay area rests conformably on the Lower Mississippian Codroy formation, but the abrupt change from fine-grained, mottled red and green sandstones of the Codroy to the coarse, red, feldspathic sandstone of the Barachois is striking. The influx of coarse red clastics signifies another sharp uplift, probably in the Labrador coast area.

No other Pennsylvanian rocks are known in Newfoundland, and hence nothing is known of the early Pennsylvanian disturbance outside the St. George Bay area.

Post-Early Pennsylvanian Phase (Appalachian Orogeny)

The major fault zone that extends from the southwestern coast of Newfoundland in a northeasterly direction to Grand Lake, White Bay, and up the east coast of the northern peninsula postdates the youngest sediments of Newfoundland. These are the Barachois series of lower or middle Pottsville (Early Pennsylvanian) age. Relief features and escarpments in other parts of the island trend northeasterly and parallel the western fault zone. These in part may also be due to faults of the same phase. Betz (1943)

suggests that the orogeny is an extension of the Appalachian orogenic belt of the Canadian and United States Appalachians.

Volcanic activity had died out by the Pennsylvanian after very little in the Mississippian, and no Carboniferous intrusions have yet been noted. The post-Barachois faulting and thrusting mark, as far as known, the last compressional deformation in the Appalachian mountain systems of Newfoundland.

Post-Appalachian History

No Triassic fault basin sediments are known as in Nova Scotia and New England, and no coastal plain sediments of Cretaceous or Tertiary age occur above water on Newfoundland. Without these signs of submergence, it is concluded that the island has been mostly above sea level since the Appalachian orogeny, and has been a site of erosion. It undoubtedly has had broad connections with the Maritime Provinces and the Gaspé in the Mesozoic. Likewise, the region of its northeastward projection into the Atlantic must have been extensively emergent in times past.

The broad banks off Newfoundland continue the continental shelf from Nova Scotia, and as late Cretaceous fossils have been dredged off Nova Scotia (see Chapter 10), one could assume the same fossil-bearing beds will be found under the Banks of Newfoundland. An enticing experiment would be the drilling of a deep well on Sable Island.

Twenhofel and MacClintock (1940) have described three fluvial erosion surfaces in Newfoundland in much the same aspects as in the central Appalachians, and hence assign a similar history of Cenozoic epeirogenic uplift to the island. The major difference is that the Maritime Provinces and Newfoundland have not emerged as much as the Appalachians south of New York City. If they should rise another 1000 feet, then much of the continental shelf would be land and probably a large bordering coastal plain with Cretaceous and Tertiary sediments would appear.

Cabot Strait Fault (?) and Seismic Profile

The *Tectonic Map of Canada* (1950) shows a fault along Cabot Strait between Nova Scotia and Newfoundland, with the implication that it is a transcurrent fault offsetting the structural elements of the two provinces.

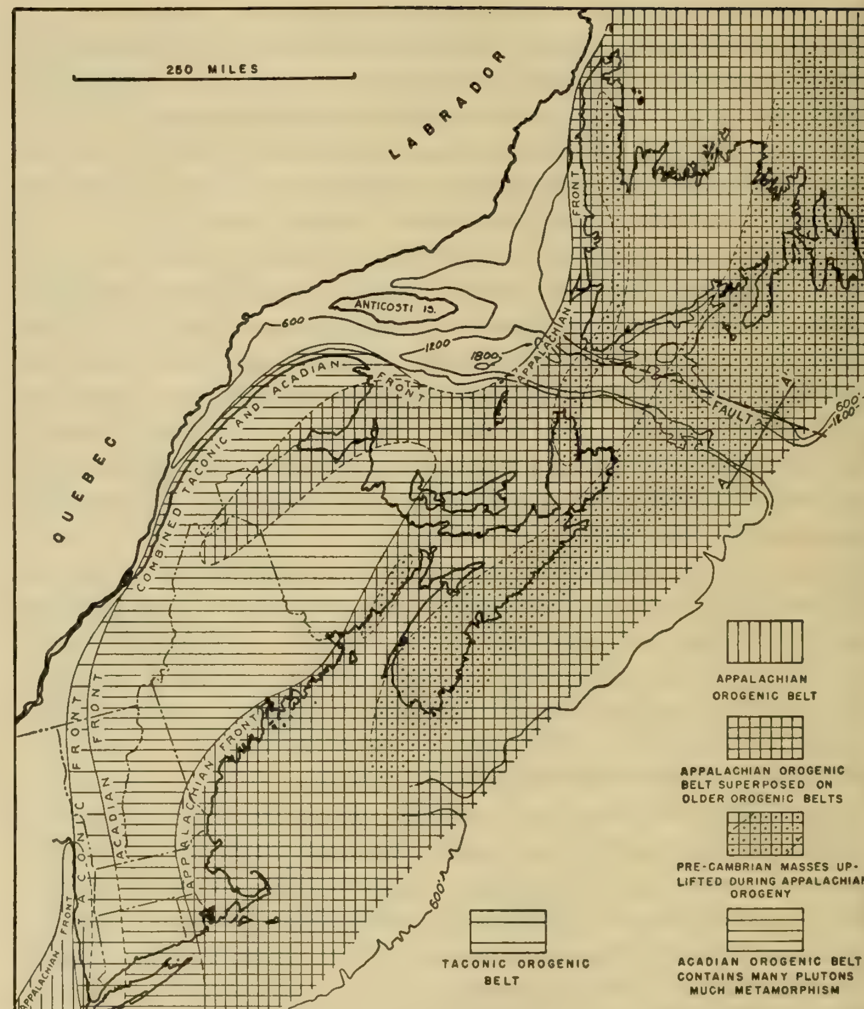


Fig. 13.12. Paleozoic orogenic belts of Greater Acadia. In addition to the Taconic, Acadian, and Appalachian orogenies there were several others in various places that are not represented. The post-Silurian Caledonian orogeny was pronounced in Newfoundland and Nova Scotia. A mid-Ordovician Vermontian is known in the Vermont-Gaspé region.

Reference to the map of Fig. 13.12 will indicate the position of the presumed fault. The structural front passes between Anticosti Island and the Gaspé Peninsula and between Labrador and the Northern Peninsula of Newfoundland under the Straight of Belle Isle (Figs. 13.1 and 13.6). Since the front is entirely submerged, its position as shown on Fig. 13.12 is only a guess. Nevertheless, the conclusion must be drawn that a deep recess in the structural front exists between the Champlain-Gaspé salient and the Newfoundland salient. Perhaps this is the result of horizontal offset along a transcurrent fault.

The submarine trough of the Gulf of St. Lawrence extends out under Cabot Straight to the edge of the continental shelf. See Fig. 13.12. It has a depth of over 600 feet for a distance of 750 miles, and from a point midway south of Anticosti Island to the shelf rim is over 1200 feet deep. At two places it is 1800 feet deep, and has a closed basin in this area about 150 miles long below the 1320-foot contour. One large tributary of the trough extends up toward the Straight of Belle Isle, and another extends along the north side of Anticosti Island.

Six seismic profiles were shot on the extensive banks off Nova Scotia and Newfoundland by Press and Beckmann (1954), and a combination of three of them across the outer end of the Cabot Straight trough is shown in Fig. 13.13. The position of the section is indicated on Fig. 13.12.

The seismic section indicates for one thing that the trough is erosional into the unconsolidated sediment layer, and this is the conclusion that Shepard (1930) reached. From a study of the shape of the submarine valley he concluded that it was first a subaerial stream valley and then was modified by glaciers flowing seaward along it. Glacial striations and roches moutonées on the southern tip of Newfoundland and on St. Paul Island off the north end of Nova Scotia demonstrate the past ice flow. The present depth of the trough is no greater than fiords elsewhere. The trough walls do not resemble fault scarps—they are straight segments with hanging valleys.

In interpreting the seismic section, Press and Beckmann say that it supports the thesis that the trough is of fault origin, yet at the same time say that the faulting occurred during the deposition of the sediments of the 3.80-km/sec layer. They regard the 3.8-km/sec layer under the north side

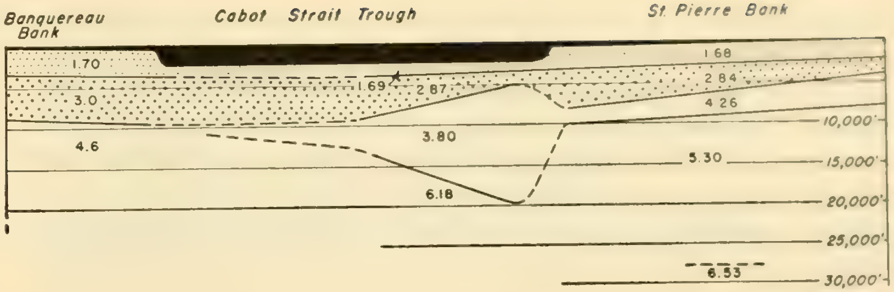


Fig. 13.13. Seismic profile across Cabot Strait, Nova Scotia, and Newfoundland. See section line A–A′, Fig. 13.12. Figures are velocities in km/sec.

of the trough (Fig. 13.13) as demonstrating the faulting. It is possible that the wedge shape of this layer does indicate faulting, but not in post-unconsolidated sediment time. The 3.80-km/sec layer is logically interpreted as consolidated sediment. Consolidated sedimentary rocks would be either Triassic red-beds or Carboniferous of the nature of the basin sediments of southwestern Newfoundland, and faults of this age are long since dead, according to the history of the Piedmont and Greater Acadia.

Mild earthquake activity is cited as evidence for the fault origin of the Cabot Strait trough. Two earthquakes whose epicenters were on the shelf slope immediately off the trough mouth have caused submarine landslides and numerous Trans-Atlantic cable breaks. Shepard questions the presumed connection between these earthquakes and continuing displacement along faults causing the trough.

Both sides of the modern trough are about the same, yet the seismic profile indicates the possibility of a fault on one side only. The conclusion is reached that in Carboniferous or Triassic time a trough formed, possibly by downfaulting, but that since then no further movement has occurred.

Now, to the original question; could the structural elements of the Maritime Provinces and Newfoundland be offset appreciably by horizontal motion along a transcurrent fault? The seismic profiles have demonstrated the possibility of a late Paleozoic or Triassic fault along the outer stretch of the Cabot Strait trough. If this fault is part of the Triassic fault system, it would probably be one of vertical displacement. If like the fault that

bounds the east side of the Carboniferous basin of the St. Georges Bay area of southwestern Newfoundland, it would also be one of vertical displacement. The wedge of sediments of the 3.80-km/sec layer suggest a vertical fault. The major structural elements from Newfoundland to Nova Scotia may be drawn across to Nova Scotia with reasonable continuity and without a horizontal offset, as shown in Fig. 13.12. Although none of these is compelling evidence against horizontal movement, they lead the writer to conclude that considerable transcurrent movement has not occurred.

MAJOR TECTONIC RELATIONS OF GREATER ACADIA

Definition

Greater Acadia has been defined by Schuchert and Dunbar (1934) as the combined regions of New England, the Maritime Provinces, the St. Lawrence–Gaspé area of Quebec, and Newfoundland. Much of the area is now covered by shallow waters, and from an historical point of view Greater Acadia includes all the lands of the past in the great geosynclinal and orogenic belt seaward to the continental shelf slope.

Major Geocynclinal Characteristics

Numerous series of beds in Greater Acadia have thicknesses of 5000 to 15,000 feet, and the total thickness in places ranges up to 100,000 feet. Thick and coarse clastics in every stratigraphic system of the Paleozoic and numerous unconformities within and between systems attest long-continued crustal unrest in the geonsyncline and at times in belts adjacent to it. A dominant lithology of the materials in the geosyncline is volcanic rocks of all descriptions. They consist chiefly of andesites and basalts, but other varieties, especially rhyolites, are by no means absent. A very thick accumulation of Ordovician rhyolite marks the central part of the geosyncline in Newfoundland. The volcanics occur as flows, in large part submarine, and as various pyroclastics. They are especially concentrated in the medial part of the geosyncline, if the Precambrian rocks of Nova Scotia and the Avalon peninsula of Newfoundland mark the site of the outer or southeastern portion. The inner belt of the Taconic Mountains–Lake Champlain–St. Lawrence–Gaspé region was comparatively free of volcanics until late Ordovician and Silurian time when the igneous activity spread to the Gaspé Peninsula and to western Newfoundland in the western belt. Aside from Devonian volcanic activity in the Gaspé Peninsula the western belt was again free of volcanism after Silurian time. Eruptive activity had died out in all Newfoundland by late Mississippian time but not in the Maritime Provinces and in the eastern part of New England. Volcanism continued exceedingly active there in places, and was accompanied and followed in the Carboniferous basins of New England by intrusive activity.

Batholiths

The central zone of the geosyncline, along with tumultuous volcanic activity, was the site of great batholithic intrusions. Where better known, as in New Hampshire, four magma series are recognized, the first about of Taconic age and the other three of Acadian, which there started in mid-Devonian and lasted probably until early Mississippian. Of the three Acadian magma series, the first preceded the major compressional orogeny, the second was synorogenic, and the third followed the orogeny.

As studies progress in the Maritime Provinces and in Newfoundland, it is becoming clearer that most of the dioritic to granitic batholiths there are Acadian also. The batholiths are not limited to the medial volcanic zone of the geosyncline but some have intruded the inner, less volcanic complement of geosynclinal sediments and others in great volume, the outer zone, now mostly of Precambrian rocks.

Metamorphism

A striking character of the stratified rocks of the geosyncline of Greater Acadia is their metamorphism. Where distant from the batholiths they are generally slates, phyllites, argillites, quartzites, and metavolcanics. Where close to the altering influence of the intrusions they are schistose and gneissic. The very-low-grade and low-grade metamorphism is more characteristic of the inner belt, and also the outer where Paleozoic sediments are preserved, as in the Conception Bay area of Newfoundland. Medium-grade metamorphism is more characteristic of the central belt.

Ultramafic Intrusions

A zone of serpentinized ultramafic intrusions extends from Georgia through the crystalline Piedmont to New York City, and from New York northward through the Taconic system to the St. Lawrence and Gaspé. From there it is believed to continue through western Newfoundland.

Fronts of Successive Orogenies

An attempt was made by Schuchert in his early paleogeographic maps and later by Schuchert and Dunbar (1934) to show the major structural elements of Greater Acadia. They postulated a western trough of sedimentation, the St. Lawrence geosyncline; a central land barrier, the New Brunswick geanticline; and eastern trough of sedimentation, the Acadian geosyncline; and beyond this, a borderland, Novascotica. As described on previous pages, the "New Brunswick geanticline" has been found to be approximately the heart of the geosyncline—a site of such sedimentation and prodigious igneous and orogenic activity. Crustal movements within the orogenic belt were numerous, and the island barriers and peninsulas were too many and transitory to be charted satisfactorily with present knowledge.

Kay (1947) has illustrated the Taconic, Acadian, and Appalachian orogenic systems of Greater Acadia to have been formed by deformation of the sediments of the eugeosyncline. This great sedimentary province includes the volcanic assemblages of sediments, the batholiths and serpentinites, in contrast to the relatively igneous-rock-free inner miogeosyncline typified by the sediments of the Ridge and Valley province. It is clear that the belts of deformation of the eugeosyncline impinge on the Canadian Shield in the Greater Acadia region, and that the belt of deformation of the inner miogeosyncline terminates approximately at the Adirondacks.

Some progress can be made toward an understanding of the spatial relations of Greater Acadia if the distribution of the orogenic belts is charted, rather than the poorly documented and transitory shore lines. The fronts of the Taconic, Acadian, and Appalachian orogenic belts are known in places with considerable precision and in others only approximately. Figure 13.12 shows these fronts, as well as the zones of superposition of one belt over the other. Evidence of the locations for the most part has already been presented, and when composed for the entire Greater Acadia, yields the picture recorded on the map. In the lower left-hand corner, the northern end of the Appalachian folded and thrust-faulted belt of the Valley and Ridge province is seen. The Taconic front then faces the shield (with its thin sedimentary veneer). At Quebec City on the St. Lawrence, the front of the Acadian orogenic belt impinges on the shield, and as far as known from Quebec City to the tip of Gaspé and beyond, the Taconic and Acadian belts are superposed. The two belts in the Gaspé Peninsula swing eastward, and even somewhat southward of east, and project in that direction into the Gulf of St. Lawrence.

Where next observable in southwestern Newfoundland, the front of the Appalachian belt faces the shield, and is impressed on all older belts. It, therefore, appears that from Vermont northeastward successively younger orogenic belts overlap inward and front on the Canadian Shield. The equivalent of the Ridge and Valley folded and thrust-faulted province does not exist north of the Catskills. In Keith's terminology the Taconic and Acadian orogenic systems compose a pronounced "salient" toward the shield in the Vermont–St. Lawrence–Gaspé region.

The map also shows linear Precambrian masses that were uplifted during the Appalachian orogeny and, if once covered by Paleozoic strata, were later subject to erosion and stripped of their mantle. The Long Range Mountains element of western Newfoundland is fairly definitely of this origin. It seems to find continuation in northern Nova Scotia, in Precambrian exposures on the western side of the Bay of Fundy, and perhaps even in Precambrian rocks in the Boston basin region. Precambrian rock forms most of the Avalon peninsula of Newfoundland and also crops out in several places west of the peninsula. It has not been proved that this region is one of late Paleozoic uplift, but only inferred because of the numerous escarpments and shore lines that parallel the known Appalachian elements of western Newfoundland, and the faults of Conception Bay which resemble those of the western Carboniferous basins. It ties in well with the extensive Precambrian area of eastern

Nova Scotia in relation to the Appalachian front, and in having a similar thick Proterozoic volcanic sequence of rocks. The zone marks the site of a great Proterozoic trough in which volcanic rocks accumulated voluminously and were frequently deformed. The Avalon peninsula contains no sedimentary rocks younger than early Ordovician and may have been an area of erosion since then. The Great Bay de l'Eau conglomerate suggests a sharp uplift of eastern Newfoundland in late Silurian or early Devonian time, and the region was probably affected by the Acadian movements and intrusions. The Precambrian of Nova Scotia contains numerous batholiths, presumably of Acadian age. It is entirely possible that the outer Precambrian uplift is one that dates back to mid-Paleozoic time and is complex.

The presence of the geanticline of Precambrian rocks along the outer exposed margin of Greater Acadia is rather significant in demonstrating that the continent has not been added to appreciably, or has not grown seaward much, since Proterozoic time.

14.

OUACHITA, MARATHON, AND COAHUILA SYSTEMS

OUACHITA SYSTEM

Location and Topography

The Ouachita Mountains occupy a belt 50 to 60 miles broad and 200 miles long in southeastern Oklahoma and western Arkansas. See maps, Figs. 14.1 and 14.2. They are somewhat like the Appalachians in topographic appearance, although not generally so high. Their level-topped subparallel east-west ridges reflect structure and dissection of erosion surfaces. The ridges rise scarcely 250 feet above the valley west of Little Rock but gradually increase in height toward the Oklahoma–Arkansas border, where the highest point is 2900 feet above sea level and nearly 2000 feet above the valley floors. Their eastern, western, and southern margins are blanketed by the Gulf Coastal Plain sediments.

Stratigraphy

The oldest rocks of the Ouachita Mountains are Cambrian, and these are exposed in the central anticlinorium. The section of the anticlinorium or "core area" of southeastern Oklahoma in McCurtain County as measured by Pitt (1955) is as follows:

Bigfork chert	?
Womble shale	66+ ft
Mazarn shale	600 ft
Crystal Mountain sandstone	50–100 ft
Collier shale	180 ft
Lukfata sandstone	150+ft

Northwestward each thrust sheet has elements of its stratigraphy, and these are given by Hendricks (1943) in Fig. 14.3.

The Arkansas novaculite is a conspicuous formation of the pre-Mississippian sequence. It has a counterpart in the Marathon uplift of west Texas, the Caballos chert, but is not present in the southern Appalachians. The Bigfork chert, Pinetop chert, and Woodford chert, as well as the siliceous nature of the limestones and shales indicate that a dominant characteristic of these formations is silica. Pitt (1955) thinks that much of the silica is secondary, having been introduced by groundwater after extensive fracturing.

The combined thickness of the Cambrian, Ordovician, Silurian, and Devonian rocks is hardly 3000 feet, and they are regarded as a shelf or platform type of deposit, although the high silica content is unusual in such a setting. The Mississippian and Pennsylvanian strata are almost entirely clastic—shale and sandstone—and are very thick. A measurement of 18,950 feet for the Ouachita Mountains sequence of Stanley, Jackfork, and Johns Valley formations is given by Cline and Moretti (1956), and 17,000 feet for the foredeep sequence of Atoka (Hendricks *et al.*, 1936).

The terms Ouachita facies and Arbuckle facies have been widely used to compare or contrast the sequences of the Ouachita Mountains and

Fig. 14.1. Composite map of the tectonic features developed in the late Paleozoic in the Mid-Continent region. Taken from R. E. King *et al.* (1942), Moore and Jewett (1942), and other publications. In Kansas the dotted names designate the older features. A.A., Arbuckle anticline.

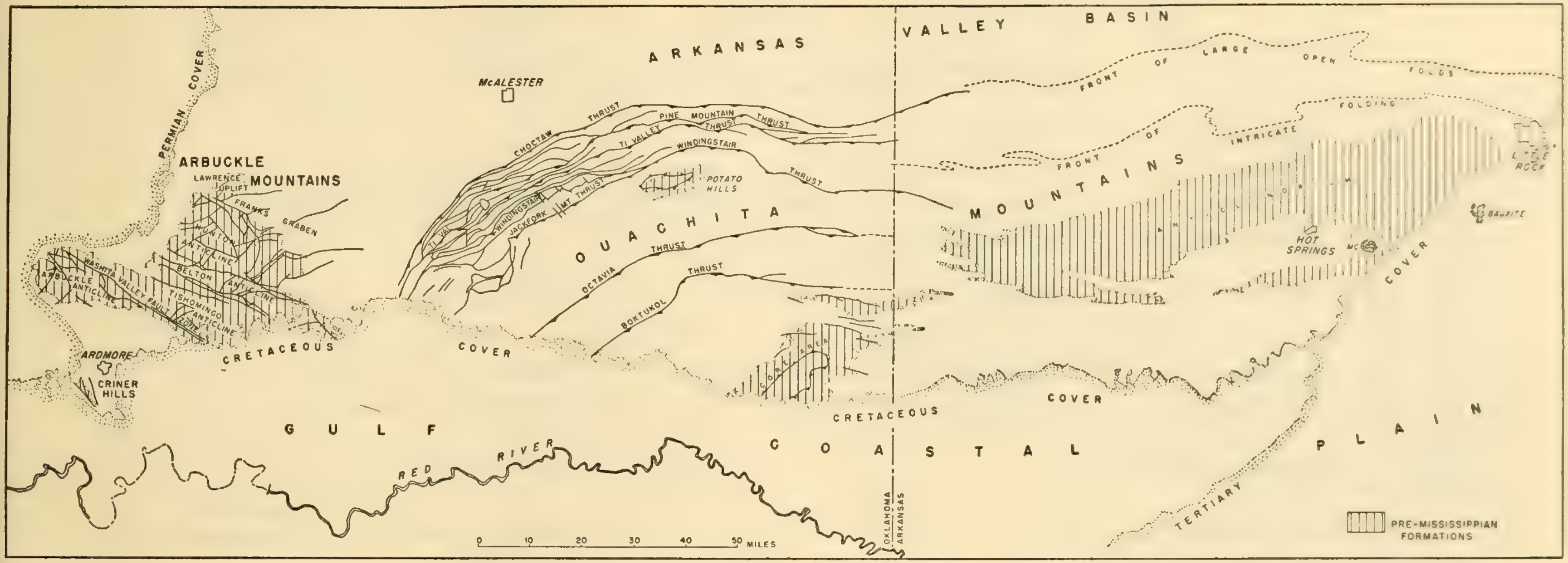


Fig. 14.2. Generalized structure map of the Ouachita and Arbuckle Mountains. MC, Magnet Cove.

the Arbuckle Mountains. The Ouachita facies is characterized by an abundance of silica in the pre-Mississippian formations and by the very thick Carboniferous clastic sequences. Also it appears that incipient metamorphism is included by some as a mark of the facies. This is all a misuse of the term facies as defined, but for local paleogeologic studies it is convenient, if properly understood.

Structure

The Ouachita Mountains may be divided into a western division, replete with thrust faults, and an eastern division, intensely folded but not appreciably faulted.

According to Miser (1929) there are five thrust sheets in the Oklahoma Ouachitas (see cross section D–D′, Figs. 14.1 and 14.4), but in light of Hendricks' additional work there are four "independent" thrusts. They are, from northwest to southeast: (1) the Choctaw fault, (2) the Pine Mountain fault, (3) the Ti Valley fault, and (4) the Windingstair fault. See Fig. 14.5. Each sheet has been thrust from south to north and has been broken by numerous smaller, high-angle reverse faults that presumably join the main thrusts at depth. Minor cross faults are numerous, and larger cross faults are present in several settings. The stratigraphy of each thrust sheet is somewhat different and is summarized in Fig. 14.3.

In front of the thrust sheets is the Arkansas Valley basin whose beds have been cast into open folds which gradually decrease in intensity toward the north. These folds partake of some of the characteristics of both its bounding provinces, the beds on the south being rather closely folded near the Ouachitas but progressively more open farther north toward the Ozark dome. Normal faults on the north side of the valley

		COAL BASIN	BLOCK S.E. of the CHOCTAW FAULT	BLOCK S.E. of the KATY CLUB FAULT	BLOCK S.E. of the PINE MTN. FAULT	BLOCK S.E. of the TI VALLEY FAULT
CARBONIFEROUS	PENNSYLVANIAN	McAlester sh. Hartshorne ss. Atoka fm. Wapanucka ls. Springer fm.	Atoka fm. Wapanucka ls. Springer fm.	Atoka fm. Chickachoc chert Springer fm.	Atoka fm. Springer fm.	Atoka fm. Johns Valley sh. Jackfork ss. Stanley sh.
CARBONIFEROUS	MISS.	Caney sh. Sycamore ls.	Caney sh.	Caney sh.	Caney sh. Sycamore ls. (?)	
DEVONIAN ?		Woodford chert			Woodford chert	
DEVONIAN	HUNTON GROUP	Bois d'Arc ls. Haragan sh.			Pinetop chert Unnamed ls.	Arkansas novaculite
SILURIAN	HUNTON GROUP	Henryhouse sh. Chimneyhill ls.				Missouri Mountain sh.
ORDOVICIAN		Sylvan sh. Fernvale ls. Viola ls. Simpson group				Polk Creek sh. Bigfork chert Womble sh.
		Arbuckle group				
CAMBRIAN		Reagan ss.				

Fig. 14.3. Sequence of strata characteristic of each of the structural blocks of the Black Knob Ridge area of the western end of the Ouachita Mountains. After Hendricks, 1943. Katy Club fault is a minor shear along the line of cross section in Fig. 14.5. The Stanley shale is now considered Upper Mississippian.

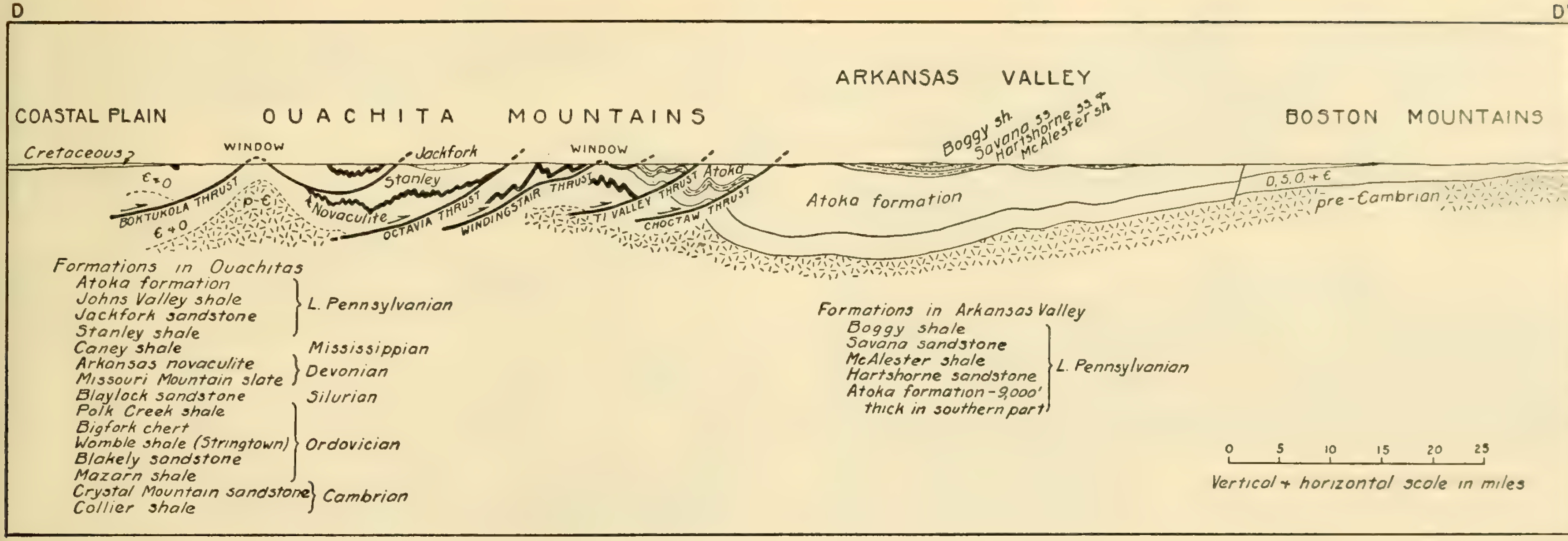


Fig. 14.4. North-south cross section through Ouachita Mountains and Arkansas Valley. Section D–D′, Fig. 14.1. Somewhat idealized from Miser, 1929, and Hendricks *et al.*, 1936.

are common (Croneis, 1930). Their south sides are generally down, thereby augmenting the basin structure.

The thrust faults appear to die out eastward into Arkansas where a fold complex indicates also considerable compression. See Fig. 14.6. An anticlinorium is the dominant structure in the approximate center of the exposed fold belt. The minor folds on the major anticline are sharp and mostly asymmetrically inclined northward. Two large anticlines with amplitudes of 7000–10,000 feet dominate the belt north of the intricately folded anticlinorium. Precambrian rock is nowhere exposed in the Ouachitas—a condition similar to that in the Valley and Ridge province of the Appalachians.

In Arkansas it is not clear just where the line should be drawn separating the folds of the Arkansas Valley basin and those of the Ouachitas. The Choctaw thrust is considered the northern boundary of the Ouachitas in Oklahoma. Numerous folds in the Arkansas Valley basin sediments are conspicuous on the *Tectonic Map of Oklahoma* (Arbenz, 1956).

The turn of the thrusts of the west end of the Ouachitas to the south is very conspicuous. The number of thrust slices increases also, and it appears that the strata were more crowded here than elsewhere. The junction with the Arbuckles is unfortunately covered by the Cretaceous sediments, but a number of wells and some geophysical work help to explain the obscure relationship. The strike of the structures and trend of the Arbuckles is nearly at right angles to the southward veering Ouachita structures, and the formations are in part conspicuously different. The problem of the relation of the Arbuckles to the Ouachitas will be taken up later.

No rocks or structural elements resembling the Blue Ridge or the

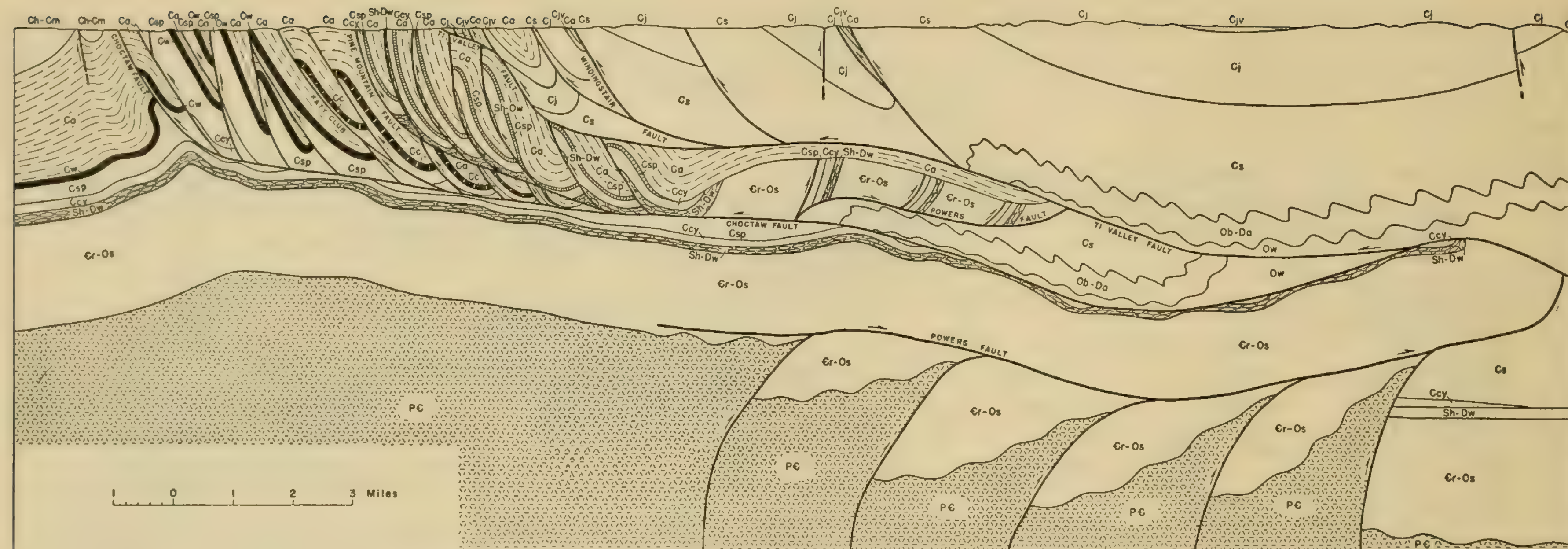


Fig. 14.5. Cross section of the Black Knob Ridge area of the western end of the Ouachita Mountains. After Hendricks, 1943. Formations may be identified by reference to chart, Fig. 14.3.

crystalline Piedmont are exposed on the south flank of the folded and thrust-faulted Ouachitas. These tectonic units have been looked for in numerous wells which have penetrated the Cretaceous and Jurassic cover, but the wells are apparently not sufficiently far enough down dip and seaward to discern the units.

Metamorphism

The pre-Mississippian formations of the central anticlinorium or "core" of the Ouachita Mountains in both Oklahoma and Arkansas are slightly metamorphosed. The shales are dynamically altered to argillites, meta-argillites, and in places to phyllites (Goldstein and Reno, 1952; Flawn, personal communication and 1956). The novaculite and chert units are most metamorphosed at the eastern end of the anticlinorium near Little Rock and at the southwestern end in McCurtain County, Oklahoma (Miser, 1943). In McCurtain County the fissility of the Cambro-Ordovician shales is parallel or subparallel with the bedding (Pitt, 1955). The small folds around the central core are overturned southward and slaty cleavage has developed which dips generally steeply north.

The position of the Ouachita front under the Cretaceous and Tertiary cover is recognized on the basis of metamorphism and high dips in contrast to the lack of metamorphism and very low dips of the beds of the foreland. See Fig. 14.6. The siliceous nature of the Devonian to Cambrian rocks of the Ouachitas is an additional guide.

Structural Problems

The *Geological Map of Oklahoma* (Miser, 1954) shows the Hendricks version of the multiple thrust structure as well as two windows, the

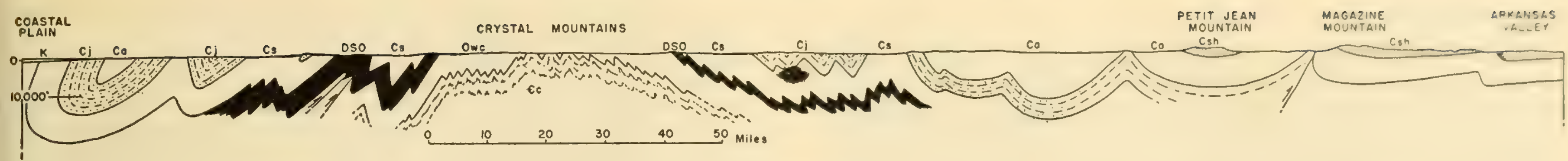


Fig. 14.6. Cross section of Ouachita Mountains in Arkansas. After cross section on *Geologic Map of Arkansas*, 1929. €c, Collier shale; Owc, Womble shale, Blakely sandstone, Mazarn shale, and Crystal Mountain sandstone; DSO, Arkansas novaculite, etc; Cs, Stanley shale; Cj, Jackfork sandstone; Ca, Atoka formation; Csh, Savanna, Paris, Fort Smith, Spadra, and Hartshorne formations.

Potato Hills and the McCurtain County core area (also called the Choctaw anticlinorium). These have been reproduced in Fig. 14.2. Hendricks' synthesis of the thrust structure involves translation of rocks considerable distances, a seeming requisite of the Ouachita overthrusting of the Arbuckles. See Figs. 14.1 and 14.6. Hendricks postulates that a deep-seated thrust plane exists, the Powers, along which rocks of "Arbuckle facies" were thrust southeastward, and then, slightly later, the strata involving the thick Carboniferous clastic sequences were thrust northward to rest as allochthonous sheets on a foreign (Arbuckle) foundation.

The *Tectonic Map of Oklahoma* (Arbenz, 1956) shows the thrust complex of the *Geologic Map* including Potato Hills window, but not the core window. The core area was remapped and reported on by Pitt in 1955, and he concluded that a normal sequence of formations exists on and around a rather simple dome—that no klippe is indicated; the previous need for a fault was due to erroneous reading of bedding and an inadequate understanding of the stratigraphic succession.

In 1957 Misch and Oles took issue with Hendricks on the basis of their own detailed mapping of the Ouachitas. They concur with Pitt on the structure of the "core" and also recognize no window in the Potato Hills. They conclude that Potato Hills is an anticlinorium of closely spaced, steep, and partly overturned folds.

The overturning is both to north and, against the direction of the supposed overthrusting, to south. Some overturned anticlinal limbs have ruptured, and steep reverse faults have developed. Some of these faults yield to the north; others yield to the south. All of these reverse faults die out along the strike, generally in the steep limbs of anticlines.

The Arkansas anticlinorium displays the same fold pattern as that seen in the Potato Hills. Steep northward and southward overturning of folds are about equal. The greatest stratigraphic and structural depth is exposed in the core of the western part of the anticlinorium (south of Mt. Ida), and there is the same continuous change in tectonic style as that found in the core of the Choctaw anticlinorium.

Misch and Oles contend that the mapped overthrusts, both major and minor, are partly steep reverse faults and partly no faults at all. The large exotic boulders of Arbuckle rocks in the Johns Valley shale are considered evidence of thrusting by Hendricks, but Misch and Oles believe they are of "deposition origin"—apparently not associated with an advancing thrust front.

Misch and Oles also believe that the differences between the "Ouachita facies" and the "Arbuckle facies" have been overemphasized.

Some units are indentical, as for example, the upper Arkansas novaculite of the Ouachitas and the Woodford chert of the Arbuckle region. Others differ relatively little, as the Bigfork "chert" and the major part of the Viola limestone, or the Stanley shale and the Caney shale. Others differ more strongly, as the Ouachita Mountains correlatives of the Simpson group. And some units differ very strongly, as the Missouri Mountain shale and the lower Hunton limestone. However, contrasted facies are not disconnected as the hypothesis of overthrusting requires. Most of the contrasted facies have transitional relationships. Some of the transitions are very gradual; others are pronounced and also have been accentuated by the intense shortening resulting from folding and faulting. None of these changes, however, exceeds those often encountered in adjacent and connecting basins, or different parts of the same basin. Moreover, the fact is often overlooked that there are marked facies changes within the Arbuckle region itself, as well as within the Ouachita Mountains.

For a review of the problems in the Ouachita Mountains see Tomlinson (1959).

Phases of Ouachita System

Early Mississippian Phase. Elevations precursory to the late Paleozoic orogeny seem to be indicated by an unconformity between the Arkansas novaculite (Devonian) and the Upper Mississippian clastics (Chaney shale). Chert conglomerates rest on the novaculite in the Potato Hills section of the Ouachitas and they are found at the base of the Stanley shale (Lower Pennsylvanian) in southern outcrops. In addition to this suggested late Mississippian disturbance, the rise in the foreland of the Ellis–Chautauqua–Ozark arch in late Devonian time may be mentioned.

Late Mississippian Phase. The deposition of more than 17,000 feet of clastic sediments of the Stanley, Jackfork, and Johns Valley formations all within a very short time indicates a great and sudden uplift nearby, which undoubtedly was one of active orogeny because a sedimentary mass of the character and quantity noted requires actively rising mountain chains. The clastics were deposited in a foredeep.

Whereas van der Gracht and others before him postulated the orogeny in the hinterland to the south, Hendricks (1943) believes that early thrust sheets came from the north and pushed southward to form a landmass. The Stanley, Jackfork, and Johns Valley shales were deposited in a basin to the south of this landmass, and the thrusting culminated in Johns Valley time. The Atoka sediments were then spread thickly over the sites of both facies. Van der Gracht believes the Atoka came from a southern highland; Hendricks does not comment on the source. The Atoka sediments reflect the second pulsation this time in the Early Pennsylvanian.

In eastern Texas, a foreland basin to the southward-trending chains of the hinterland came into existence, and in the basin the Strawn and Millsap formations were deposited, having been derived from an eastern source.

Mid-Pennsylvanian (?) Phase. The age of the major deformation of the Ouachitas is believed by several authors to have occurred in post-Atoka and pre-Boggy time. According to Fitts (1950);

> The unconformity at the base of the Boggy formation is the largest within the Pennsylvanian of Oklahoma and is probably the most widespread. Along the line of outcrop, it is progressively underlain by Pennsylvanian beds from Savanna to Atoka, locally in the Tri-State area upon Mississippian and in western areas of Oklahoma all formations down to granite.
>
> The top of the Boggy is marked by another unconformity, this one of more importance locally and to the westward in the Seminole region. The section of beds above this unconformity is generally devoid of any angular discordance and for the first time can be seen a relationship which will persist through the rest of the Pennsylvanian and lower Permian; i.e., predominantly limestone in the north grading to shales and clastics in the central to coarser clastics and red beds as the Arbuckle Mountains are approached.

The deformation of the Arbuckles in the Mid-Pennsylvanian influenced the development of the red-bed facies in the upper Cisco and Lower Permian, but later in Permian time much clastic material in the Wichita system came from an eastern source (Cheney, 1929).

Drilling operations have penetrated a formation, the Morehouse, under the coastal plain sediments, in northern Louisiana, which contains "late Paleozoic fossils" (Imlay and Williams, 1942). Its areal relations have been worked out for a limited distance in southern Arkansas and also, to some extent, its stratigraphic relations (Philpott and Hazzard, 1949; Fisher *et al.*, 1949). See Fig. 14.6. It occurs above the Eagle Mills formation and below the Louann salt and Werner formation. (Philpott and Hazzard, 1949). According to the usage of Imlay and Williams, the Louann salt and Werner formation make up the Eagle Mills. At any rate, the Eagle Mills seems to overlie the folded Ouachita facies unconformably, and if such is the case, the Ouachita thrusting predates the Eagle Mills and Morehouse. When their age eventually is fixed, the age of the Ouachita thrusting possibly will be fixed more definitely than is now possible.

Connection of Ouachitas and Appalachians

Spatial Relations. The relation of the Ouachita system to the Appalachian is hidden by the Cretaceous and Tertiary rocks of the Mississippi embayment, but they have been traced by deep wells to within 60 miles of each other. See map, Fig. 14.7. Both are strongly folded and faulted, and in both there has been thrusting toward the central stable region of the continent. In both areas there is a thick development of Early Pennsyl-

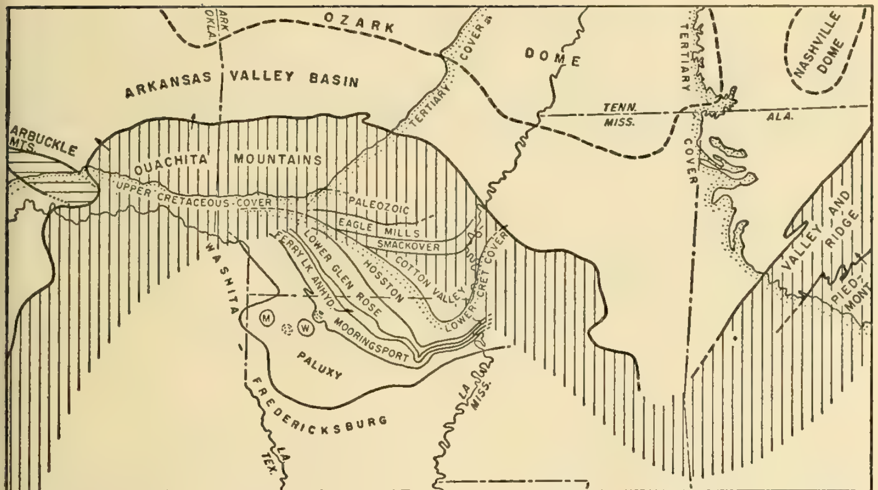


Fig. 14.7. Relation of Ouachita Mountains to southern Appalachians under the Coastal Plain cover. The pre-Upper Cretaceous geology of Arkansas and Louisiana is by Fisher, Kirkland, and Burroughs (1949). Fredericksburg, Paluxy, Mooringsport, Ferry Lake anhydrite, Lower Glen Rose, and Hosston formations are Lower Cretaceous; the Smackover and Cotton Valley are Upper Jurassic; the Eagle Mills is possibly Lower Jurassic (King, 1950a) or Permian (Philpott and Hazzard, 1949).

vanian clastic rocks derived from the hinterlands. See paleotectonic maps, Plates 5 and 6.

According to King (1950a):

The sequence of Paleozoic deposits in the Ouachita Mountains resembles that in the Valley and Ridge province in that it is composite, the older part indicating quiet deposition, and the younger part deposition during a time of considerable crustal mobility. It differs in that the boundary between the older and younger parts is post-Devonian rather than Middle Ordovician as in the Valley and Ridge province, so that there is no representation of the Taconian orogeny. Moreover, the deposits of the older part are black graptolite shales, bedded cherts, novaculites, and fine sandstones, rather than carbonates, and hence are of "eugeosynclinal" facies, as contrasted with the "miogeosynclinal" facies of the Valley and Ridge province to the east, and of the Arbuckle and Wichita Mountains farther west in Oklahoma. Deposits of the younger part, laid down under conditions of greater crustal mobility, are of early Pennsylvanian (Springer) age, and probably formed in response to the Wichita period of orogeny. They are similar to the thick late Mississippian and early Pennsylvanian deposits of the Valley and Ridge province in Alabama. The deposits of the Ouachita geosynclinal were remarkably persistent in character, for nearly the same units are present in the extension of the system in the Marathon region, Texas, many hundreds of miles to the southwest.

The Appalachian folds have been traced as far southwest as Marengo County, Alabama, on line of strike from the exposed structures of the Valley and Ridge province in the Birmingham district, where wells have encountered Ordovician limestones and dolomites directly beneath the Mesozoic.

The Ouachita folds have been traced southeastward from their outcrops in the Ouachita Mountains, across the Mississippi Embayment and into central Mississippi. Here, the boundary between Paleozoic rocks of Ouachita facies and the foreland rock trends southeastward. That this is likewise the strike of the folding is suggested by the fact that folds in the adjacent Black Warrior Basin trend southeast. In Newton and Neshoba Counties, Mississippi, near the boundary between the Ouachita area and the foreland, wells have encountered Ordovician limestones and dolomites below the Mesozoic. These are of Appalachian or Arbuckle facies, rather than Ouachita facies, which indicates the existence of an intermediate slice between the Ouachita folds and the foreland.

The Appalachian and Ouachita systems have thus been traced by drilling to within about 60 miles of each other, and they seem to be approaching at an acute angle. Southward, they pass beneath the thick Jurassic and Lower Cretaceous deposits of the Gulf Coastal Plain, so that their point of junction is beyond the reach of the drill.

Connection of Ouachitas and Marathons

The Ouachita thrust sheets not only overlie the east end of the Wichita system, but continue southward under the Cretaceous rocks of the Gulf Coastal Plain. If not the thrust sheets, the deformed strata of the orogenic belt wrap around the Llano uplift of Texas and connect with the Marathon Mountains to the west. Miser and Sellards (1931) have traced the Ouachita front under the Cretaceous strata by means of well records south to the Llano uplift, and Sellards (1931) has traced the geosynclinal rocks westward from the uplift to the Marathon exposures, also by means of well records. Flawn (personal communication and 1956) more recently has mapped this front in considerable detail.

MARATHON SYSTEM

Location and Principal Structures

Paleozoic formations appear in the Marathon basin of trans-Pecos Texas, and there reveal another great orogenic system. The Marathon

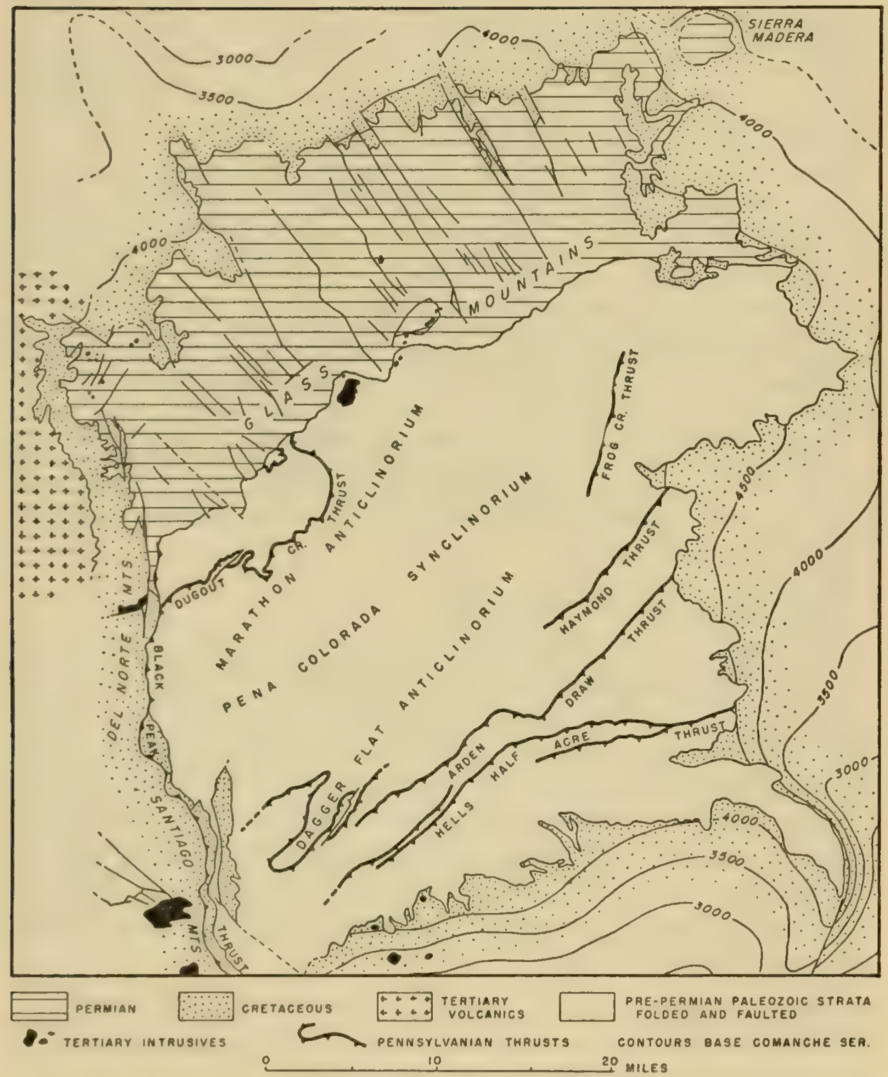


Fig. 14.8. Structure map of the Marathon uplift. After King, 1937. Black Peak thrust is post-Cretaceous. A number of Cretaceous outliers in the Paleozoic area not shown.

region lies on the edge of the Mexican highlands physiographic province which merges with the Great Plains on the east. Structurally, the region is a broad dome of Cretaceous rocks, from whose central part the Cretaceous cover has been stripped away, leaving an area of low country in the center, the Marathon basin. See Figs. 14.1 and 14.8 and cross-section M–M′–M″, Fig. 14.9. Here, strongly folded Paleozoic rocks are exposed. The Paleozoic rocks in the basin, and in the Glass Mountains which flank it on the northwest, have a thickness of 21,000 feet. The greater part of them was laid down in a subsiding trough commonly referred to as the Llanorian geosyncline. The oldest rocks are Upper Cambrian sandstones and shales, whose base is not exposed. Overlying them are 2000 feet of Ordovician rocks composed of shaly limestone and shale, with some beds of chert. The Ordovician is overlain by the Caballos novaculite, possibly of Devonian age, which reaches 600 feet in thickness. The Caballos novaculite is over-lain by a great series of clastic rocks of Pennsylvanian age, as much as 12,000 feet thick in the southeastern part of the area but much thinner in the northwest.

Llanoria and the Llanorian Geosyncline

The belt of folded sedimentary rocks of the Ouachita Mountains extends around the Llano uplift to the Marathon region and thence southwestward across the Solitario near the Rio Grande and on into Mexico. See Plate 8. The early Paleozoic trough lay about 100 miles north of the present mountains (Barton, 1945). See Fig. 14.10. In Permian time, a trough of geosynclinal proportions existed in Coahuila, 200 miles south of the Solitario.

Pre-Carboniferous sediments of the Marathon and Solitario uplifts, like those of the Ouachitas, are rather thin and include much clastic material. They are composed of sandstones, conglomerates, boulder beds of debated origin, and impure limestone with much shale, chert, and, conspicuously, novaculite. Some of the sediments evidently accumulated at no great distance from shore; others such as the shales may have been carried much farther away from their source. In the foreland areas of both the Ouachitas and Marathons, the sediments are mostly limestones. It is generally concluded that the early Paleozoic sediments came from a

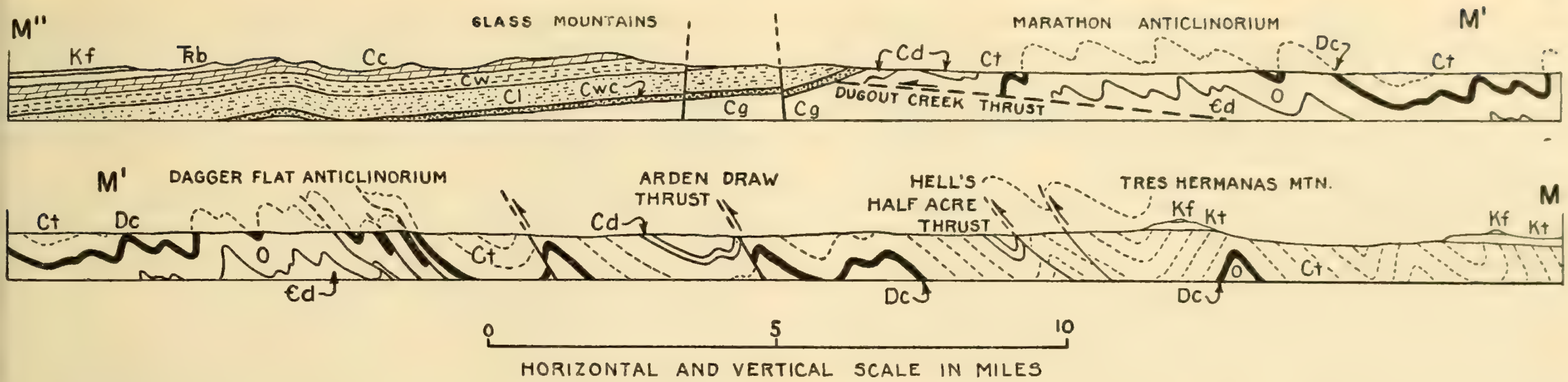


Fig. 14.9. Cross section of Marathon uplift and Permian basin. Taken from King (1937, Pl. 23, section B–B′). Ɛd, Dogger Flat sandstone (Cambrian); O, Maravillas chert, Woods Hollow shale, Fort Peña formation, Alsate shale, and Marathon limestone (Ordovician); De, Caballos novaculite (Devonian ?); Ct, Tesnus formation, Cd, Dimple limestone, Ch, Hamond formation, and Cg, Gaptank formation (Pennsylvanian); Cwc, Wolfcamp formation, Cl, Leonard formation, Cw, Wood formation, and Cc, Capitan limestone (Permian); Ћb, Besset conglomerate (Triassic ?); Kt, Trinity group, and Kf, Fredericksburg group (Lower Cretaceous).

landmass to the south or southeast and accumulated in a sea whose shore lines moved back and forth, but the propriety of calling the basin of deposition of that time a geosyncline with only 1500 to 3100 feet of sediment has been questioned (Sellards and Baker, 1934). Deep wells have enabled Barton (1945) to diagram the extent of the pre-Pennsylvanian deposits with more detail than heretofore. See Fig. 14.10. He shows that the axis of the basin was considerably north of the later Pennsylvanian, and also that the basin was too shallow to deserve the name geosyncline.

In both of the regions, the deposits of Carboniferous time attained a great thickness, possibly over 20,000 feet in the Ouachitas and 12,000 or

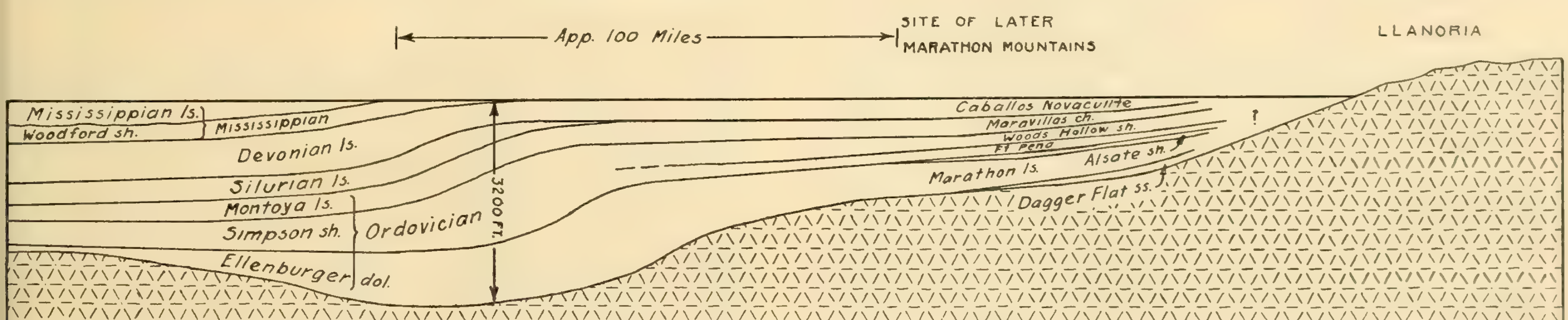


Fig. 14.10. Pre-Pennsylvanian basin of deposition in the region of the Marathon Mountains, after Barton, 1945. The section extends approximately north-south through the Marathons and into the Delaware basin, and restores the strata diagrammatically to their pre-Pennsylvanian condition.

more feet in the Marathons. The trough in which these Carboniferous sediments accumulated appears to have extended uninterrupted from the Ouachita to the Marathon and Solitario regions. This Pennsylvanian trough is referred to as the Llanorian geosyncline. The Fort Worth (Strawn) and Kerr basins seem to be expansions of the geosyncline over the margin of the foreland.

The land area of Llanoria, southeast of the Llanorian geosyncline, appears to have been composed largely of crystalline rocks and probably stood as a highland or mountain area during a large part of Paleozoic time. For the most part, the former highland is now buried beneath Cretaceous and younger strata, and the hypothesis of its former existence is based largely on evidence supplied by the composition of the Paleozoic sediments in the geosyncline (Miser, 1929; King, 1937).

Both Pennsylvanian clastic and Devonian cherty formations thicken southeastward across the Llanorian geosyncline in the Marathons; limestones are replaced by shales or cherts; and the clastic deposits contain grains of schistose or granitic rocks, pebbles of vein quartz, and cobbles of igneous rocks. The distance south at which the land lay during Paleozoic time is unknown, but it may have been 100 or more miles away. Examine Fig. 14.10.

Phases of Marathon System

Early Pennsylvanian Phase. The lowest of the Pennsylvanian formations, the Tesnus, was deposited in the Llanorian geosyncline, probably in early Pennsylvanian time (King, 1937). It is a great mass of interbedded sandstone and shale in thin and thick beds, nearly barren of fossils. In the southeastern part of the basin it exceeds 6500 feet in thickness, and it is predominantly sandstone with many arkose layers and several prominent massive layers of white quartzite. In the northwestern part of the basin, it is about 300 feet thick and is nearly all black shale with a few sandstone beds. The Tesnus, the Dimple limestone, and the lower part of the Haymond formation make up the flysch facies—a European term to signify sediments deposited during the time of a rising hinterland and a sinking geosyncline. The Dimple limestone is over 1000 feet thick in the Marathon basin, and thins southward. The Haymond formation is a mass of arkosic sandstones and shales 3000 feet thick.

Overthrusting in the southern part of the Marathon area began at this stage, as is suggested by a remarkable layer of mudstone in the upper part of the Haymond, in which are embedded large blocks of older rocks. The blocks are believed to have been derived from the erosion of advancing thrust sheets and to mark the first strong compression in the region (King, 1937).

Late Pennsylvanian Phase. The uppermost Pennsylvanian formation, the Gaptank (Upper Pennsylvanian in age), consists of conglomerate and sandstone derived from the erosion of rising folds. The strong deformation to which the Paleozoic rocks of the Marathon basin have been subjected apparently culminated after the deposition of this Upper Pennsylvanian formation. The Permian rocks of the Glass Mountains to the northwest rest, at least in places, with great angular unconformity on the disturbed older beds. See section M–M′–M″, Fig. 14.9. The structural features consist of close folds that trend northeast and are overturned to the northwest, and several thrust faults. The faulting culminated on the northwest in the nearly flat-lying Dugout overthrust, with a known displacement of more than six miles. Farther southeast the other thrusts have miles of displacement and some are folded and therefore older than the frontal fault (P. B. King, 1937).

COAHUILA SYSTEM

Known Geologic History

Exposures of Late Pennsylvanian (?) and Permian rock in the southwestern part of the Mexican state of Coahuila, some 250 miles south of the Marathon region of Texas, are believed to reveal a continuation of the Llanorian geosyncline and the approximate position of the west margin of Llanoria. In the Acatita–Las Delicias area, according to Kelly (1936) and R. E. King *et al.* (1944), a series of sediments and interstratified igneous rocks over 10,000 feet thick was deposited in a subsiding trough. The sediments came from the landmass of Llanoria, and the lava flows, sills, fragmental igneous material, and graywacke came from the west. The volcanics are rhyolite, andesite, and basalt flows and tuffs.

Late Pennsylvanian (?) limestones, possibly in part of reef origin, were deposited simultaneously with products of volcanic activity. Coarse

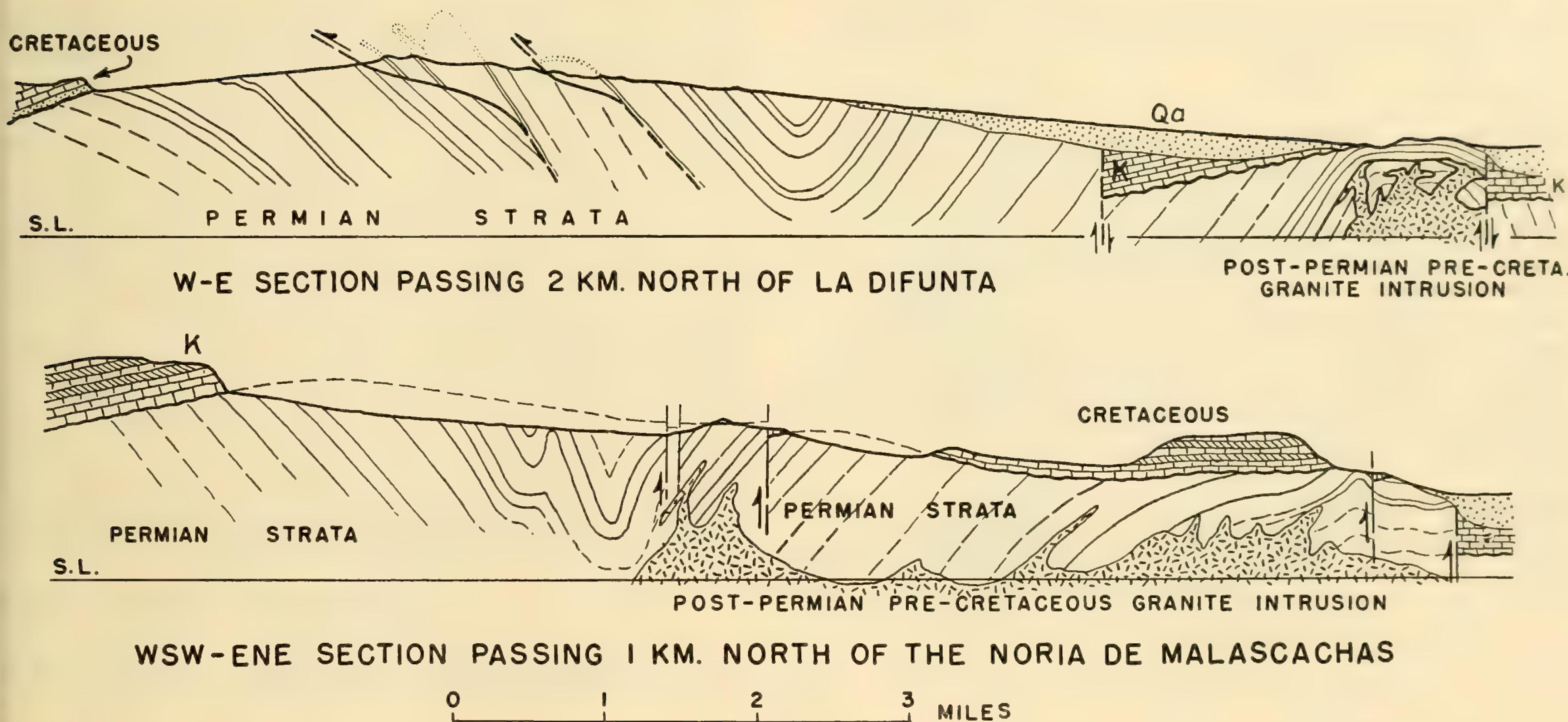


Fig. 14.11. Cross sections near Las Delicias, Coahuila, Mexico. The Permian strata consist of interbedded conglomerate, graywacke, sandstone, shale, limestone, and intermediate and basic lavas. The graywacke and lava make up about 60 percent of the sequence. After R. E. King *et al.*, 1944.

detritus from these and older rocks accumulated in the western part of the area either contemporaneously with the reefs or as a clastic wedge on the flank of an early Permian uplift. The coarseness and unsorted character of the boulder conglomerates indicate that the boulders must have been transported by unusual processes. During the remainder of Permian time, the geosyncline received deposits of clay from Llanoria on the east, flows of lava from fissures in the basin to the west, and volcanic detritus derived from the reworking of pyroclastic deposits and possibly by action of waves on the lava flows.

At some time beween Late Permian and Late Jurassic, the Pennsylvanian (?) and Permian rocks were intensely folded and overthrust. See cross section, Fig. 14.11. If the deformation took place in Late Permian time, it was the last phase of orogeny affecting the sediments of the Paleozoic geosyncline. Possibly it occurred in Early Jurassic or Triassic time, but not as late as the Nevadan disturbance (Late Jurassic), because the Upper Jurassic Oxfordian sediments rest unconformably upon the truncated Permian. The geosyncline is shown as deformed in Late Permian time on the tectonic map of Plate 8. See Chapter 17 and Fig. 17.9.

The folded rocks were intruded by batholiths of granite and granodiorite before Oxfordian time.

Structural Trends

The dominant strike of the beds in the northern part of the main Permian area is N. 35° to N. 50° E. The strike of secondary cleavage is N. 75° E. and may indicate the trend of the axis of the folds. In the southern part of the Permian area the strike of the beds swings sharply to S. 40° E. R. E. King *et al.* (1944) suggest that this may mean that the Las Delicias area is a salient part of a mountain arc in the Paleozoic structure which possibly controlled the outline of the Coahuila peninsula of Upper Jurassic and Lower Cretaceous time. Post-Cretaceous folds in continuation of this S. 40° E. trend may have been controlled by Permian folds, and thus indicate the trend of the older structures. King's suggestion is illustrated on the tectonic map.

Relation to Marathon System

On previous pages it has been explained that the folding and thrusting in the Marathons reached a climax in late Pennsylvanian time. See Plate 7. Thereafter the compressed structures were deformed only by epeirogenic uplift. The Pennsylvanian and older rocks were deeply eroded, and eventually Permian deposits overlapped them progressively southward. The tectonic map of the Permian shows the previously deformed belt as one of epeirogenic uplift. The Permian Delaware and Marfa basins were continuous with the Coahuila Permian basin; but while saline residues were being deposited in the northern basins, waters of normal salinity persisted in the south basin and probably replenished the evaporating waters to the north. After the Permian deposition in both the north and south basins, the folding and intrusions of the Coahuila area occurred.

The Coahuila structures have commonly been tied to the Marathon belt, which lies 250 miles to the north. The Permian volcanics and the post-folding granitic intrusions present characteristics foreign to the Ouachitas and Marathons in late Paleozoic time, and the writer is inclined to favor a connection with the early Nevadan belt of western Nevada and California, where the same characteristics hold. This correlation, however, presents problems in working out logical map relations.

15.

WICHITA AND ANCESTRAL ROCKIES SYSTEMS AND THE TEXAS FORELAND

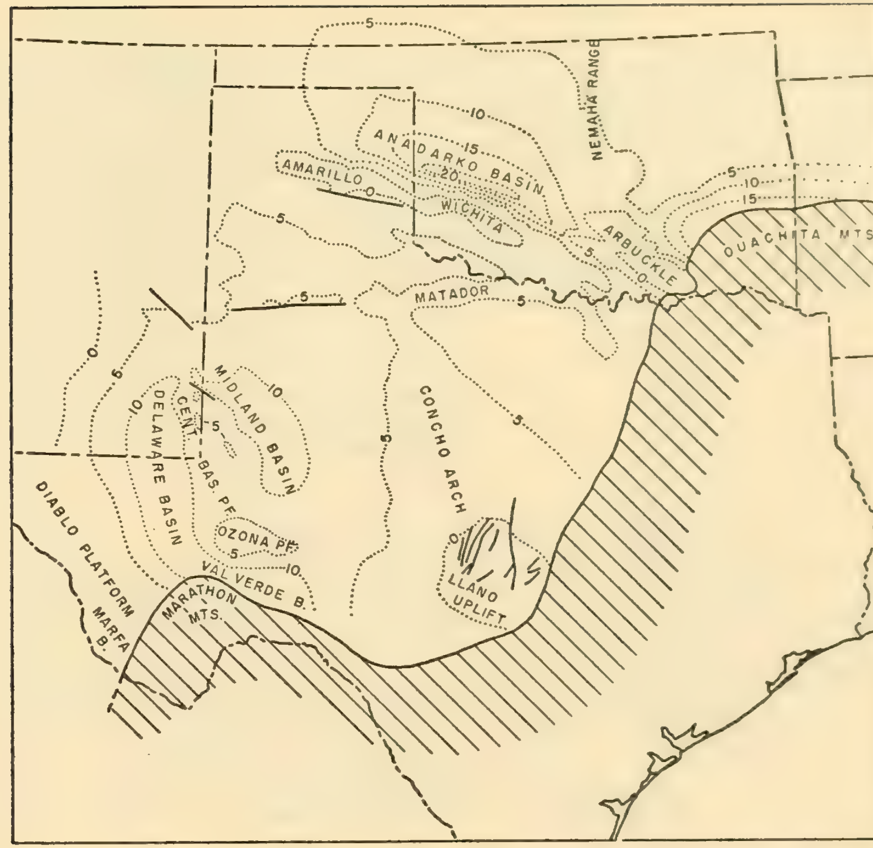


Fig. 15.1. Configuration of the Precambrian surface in Oklahoma and Texas. After Flawn (1956) and others. Numbers on contours are in thousands of feet below sea level.

WICHITA SYSTEM

Ranges and Basins of the System

Wichita Mountains. The Wichita Mountains in southwestern Oklahoma rise 1100 feet above the plains and 2480 feet above sea level. The hills are chiefly granite surrounded and embayed with nearly horizontal Permian strata. Outcrops of Arbuckle limestone of the same facies as in the Arbuckle Mountains are numerous, especially along the north side; and others on the south side and within the hills indicate that three *en echelon* anticlines are present, with granite in the cores, Arbuckle limestone on the flanks, and both overlapped unconformably by the Permian strata. The relief of the buried Precambrian surface is shown in Fig. 15.1, and a cross section at the eastern end of the range reveals the structure (Fig. 15.2). Intricate folding is described by Hayes (1952).

Amarillo Range. The Amarillo Range in the Texas Panhandle is a series of buried hills without surface expression, and is cored by Precambrian crystalline rocks. The buried hills are known to extend 125

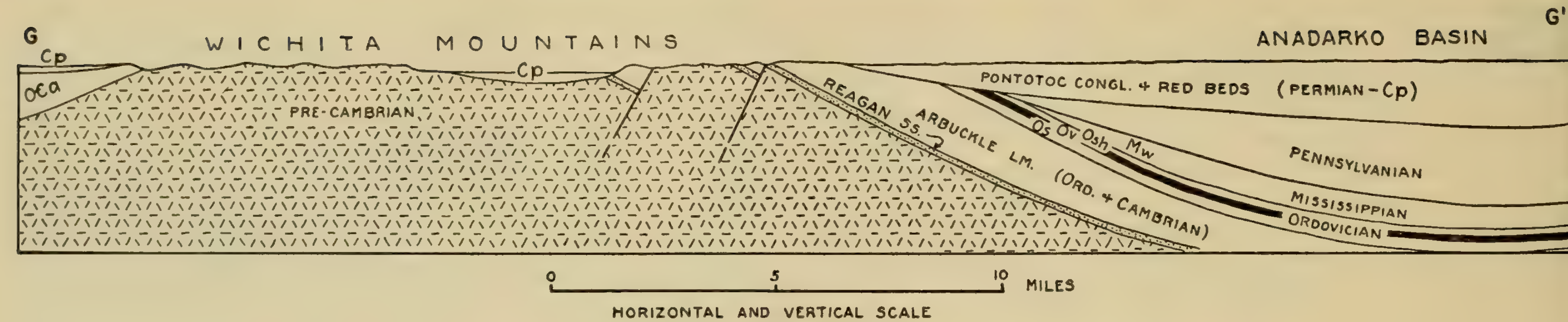


Fig. 15.2. Section through Wichita Mountains and Anadarko basin. Compiled from Taff (1904) and Millison and Reed (1939). Os, Simpson group; Dv, Viola limestone; Osh, Silvan and Hunton formations; Mw, Woodford formation.

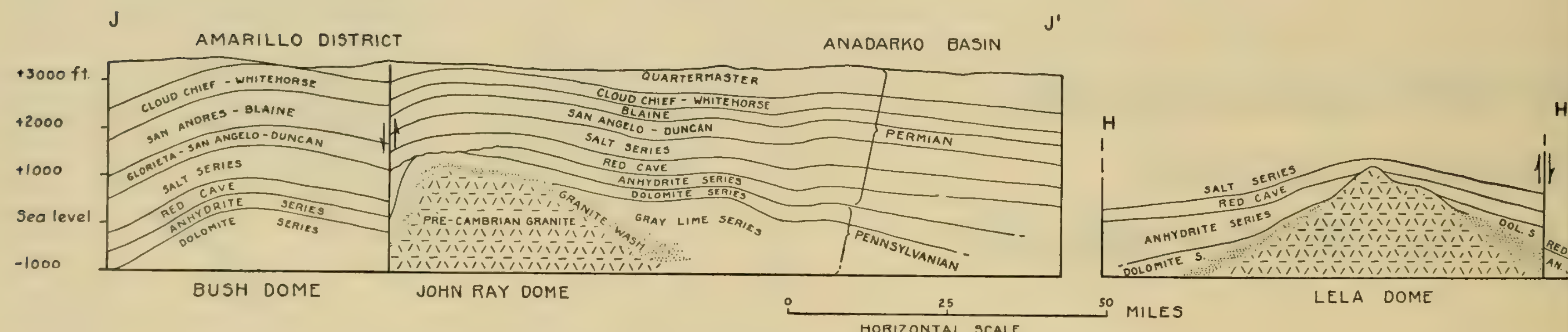


Fig. 15.3. Sections J–J′ and H–H′ of Fig. 14.1 across the buried Amarillo Range. Taken from Cotner and Crum, 1933.

miles east-west across the Panhandle. See Fig. 15.1. The highest peaks reached by the drill are about 1300 feet above sea level and 2000 feet below the surface. Some of the granite peaks are overlain directly by the Permian beds, but others are covered with the Pennsylvanian. See cross sections, Fig. 15.3. *En echelon* faults bound some hills and help produce ridges in *en echelon* arrangement. The Armillo Range and the Wichitas are continuous as shown in Fig. 15.4.

Las Animas Arch. The Amarillo Range probably extended to southeastern Colorado and northeastern New Mexico, where it joined other ranges and an arch known as the Las Animas (Maher, 1946). See the tectonic map of the Early Pennsylvanian and Fig. 15.5. The Precambrian rocks may have been exposed above sea level in Early and Mid-Pennsylvanian time along the Las Animas arch, but the thinning of the Pennsylvanian and Permian strata over the arch is chiefly due to subsidence of

the crust on either side at a more rapid rate than the arch itself. A structural relief of 3000 to 4000 feet appears to have formed during these times. Still further arching occurred in post-Paleozoic time, accentuating the structural relief.

Muenster anticline. The Muenster anticline or arch is the southeastern end of the Amarillo Wichita uplift. See Fig. 15.4. Like the Amarillo Range it is completely buried and was rangelike at the time of uplift, during the Pennsylvanian. Altogether the Amarillo–Wichita–Muenster alignment makes up an uplift with a Precambrian core and flanking truncated Lower and Middle Paleozoic strata 350 miles long.

Criner Hills. The Criner Hills are a complexly faulted horst consisting largely of Arbuckle limestone which is exposed at the surface and is flanked by Pennsylvanian and Permian strata. The horst is the east end of an anticline off the Amarillo–Wichita uplift. See Figs. 15.4 and 15.6.

Matador Arch. The Matador arch as here defined is made up of a narrow series of east-west-trending buried granite hills which extend from the New Mexico line across the Llano Estacado to Wichita Falls and beyond, a length of some 300 miles. If the overlying Cretaceous and late Paleozoic deposits were removed, the uplift would be found to consist of scattered peaks rising above an upland. Strong faults and folds trend obliquely across the uplift in a northwest direction, and these have produced an *en echelon* character to the topography (the buried peaks) and to the "highs" of the overlying formations. The Upper Pennsylvanian rests directly on the Precambrian in some localities.

Parts or all of the Matador arch have variously been called the Red River uplift, the Electra arch, and the Matador arch. The term Matador arch appears to be gaining general acceptance. The string of small uplifts produced islands in the Pennsylvanian seas and because of this the feature has also been called the Matador archipelago.

Palo Duro and Hardeman Basins. The general depression between the Amarillo–Wichita uplift and the Matador arch is filled with Pennsylvanian and Permian sediments, and has a western and an eastern division, as may be seen on Fig. 15.4. The western is the Palo Duro basin and the eastern the Hardeman basin. Various names have been used for the

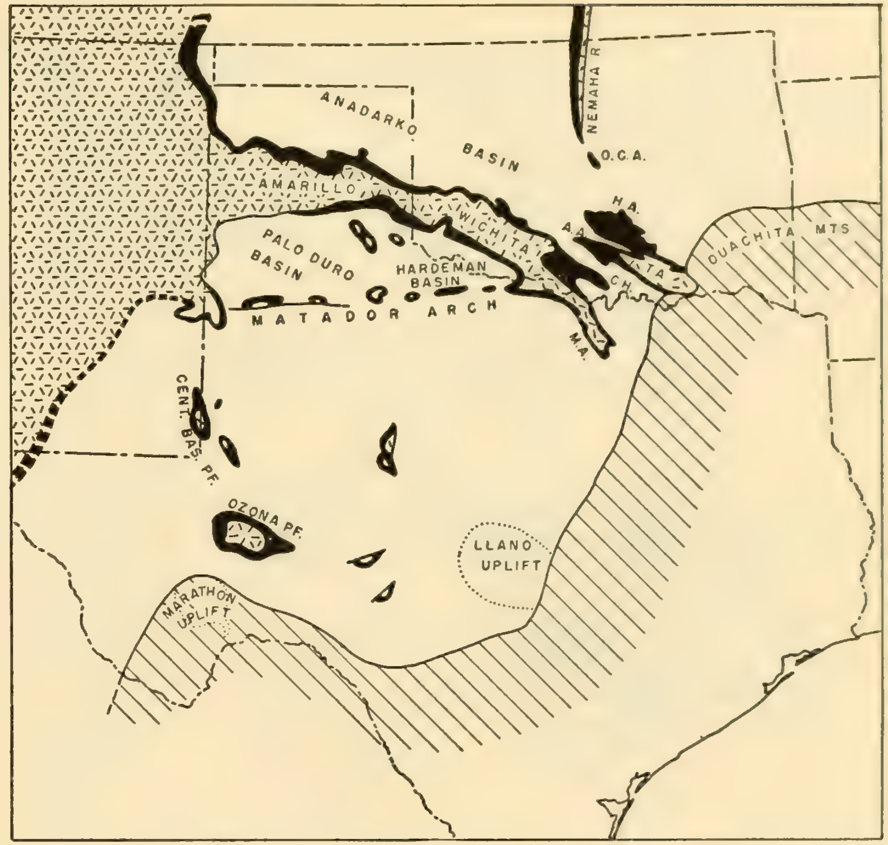


Fig. 15.4. Generalized paleogeologic map, Texas and Oklahoma, of pre-Pennsylvanian rocks. Black is sub-Pennsylvanian and Permian outcrop of Cambrian, Ordovician, Silurian, and Devonian formation. Hachured area is Precambrian. Mississippian outcrops not shown. After Totten (1956), Flawn (1956), and others. The Pennsylvanian and Permian cover has been eroded away in places in the Wichita and Arbuckle and in the Llano uplift. Doming of the Marathon uplift is post-Cretaceous. H.A., Hunton arch; A.A., Arbuckle anticline; C.H., Criner Hills anticline; M.A., Muenster anticline; Cent. Bas. Pf., Central Basin platform; O.C.A., Oklahoma City anticline.

features of this region as drilling has progressed and the geology become better understood.

Arbuckle Mountains. Topographically the Arbuckle Mountains are the hills between Davis and Ardmore, and are the surface expression of

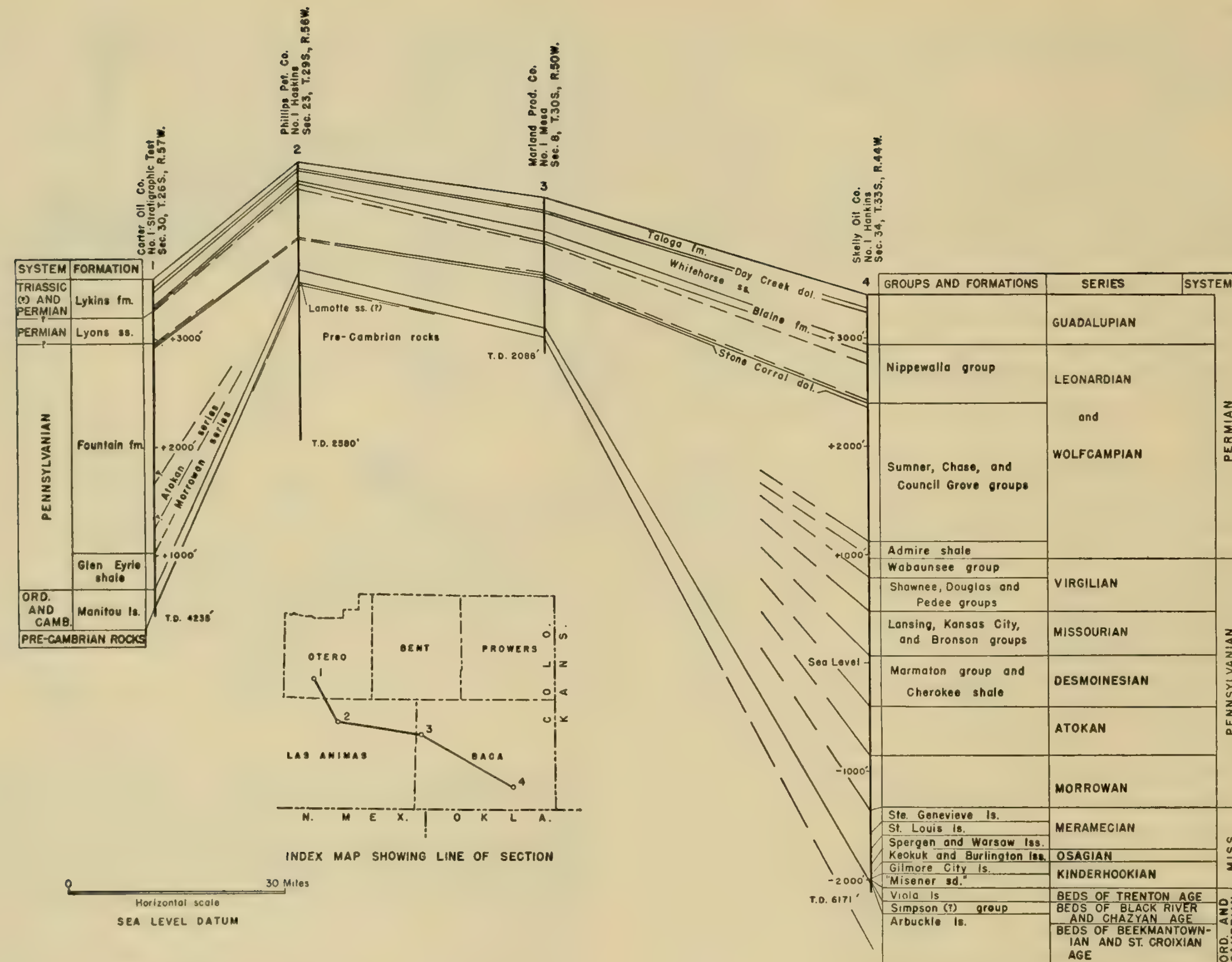


Fig. 15.5. Correlation of Paleozoic rocks across the Las Animas arch in Baca, Las Animas, and Otero Counties, Colorado. Stratigraphic classification on right is mainly after State Geological Survey of Kansas; that on left follows common usage in Colorado. From Maher, 1946.

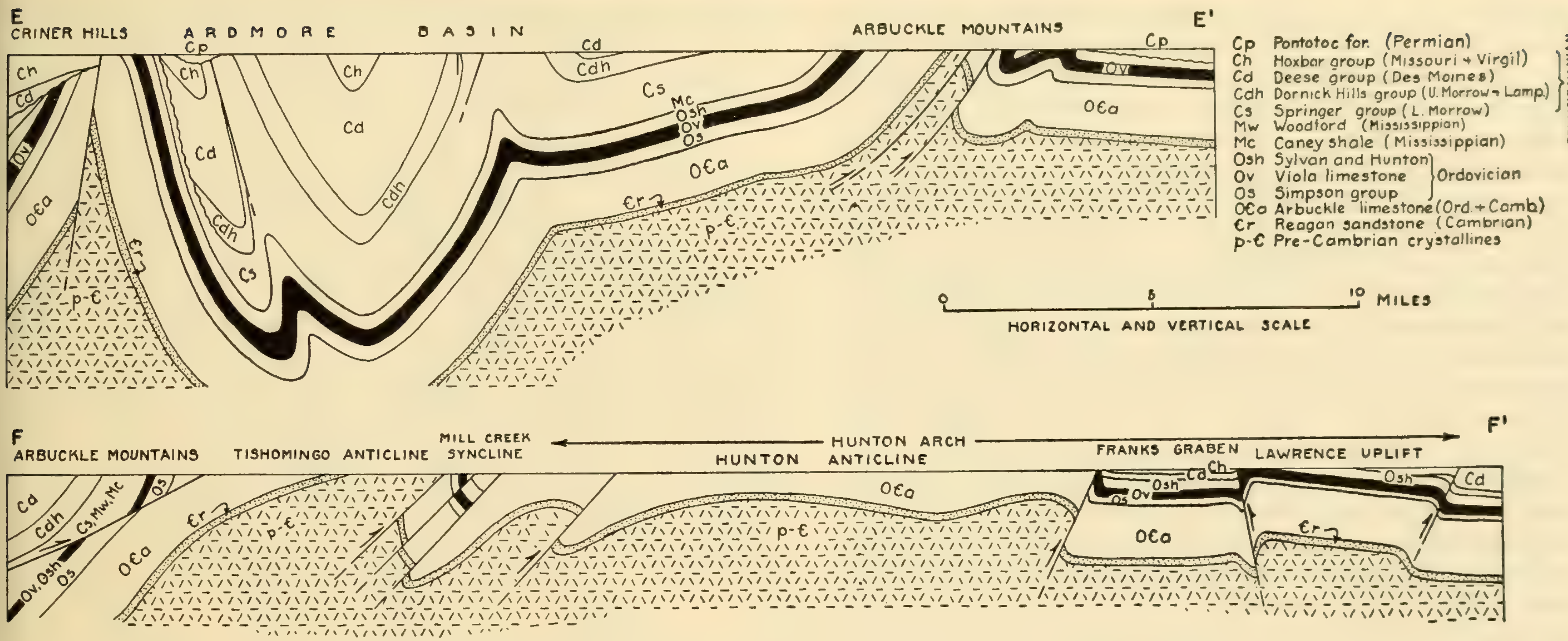


Fig. 15.6. Cross sections through the Ardmore basin, Arbuckle Mountains, and Hunton arch, compiled from Dott (1934), Tomlinson (1929), and Moore *et al.* (1944).

a large, complex anticline. In the core of the anticline, two prominent peaks of Precambrian porphyry (the Timbered Hills) rise 700 feet above the valley of the Washita River and 1400 feet above sea level. Geologically the term Arbuckle Mountians applies also to the hilly area to the north and east in which lower Paleozoic rocks crop out and where structural features of mountain proportions are located. A thick sequence of rocks from Precambrian to Late Pennsylvanian is exposed in the range. See cross sections, Fig. 15.6.

The regional structure of the Arbuckle Mountains is a series of much-faulted subparallel folds trending northwest and southeast. They are shown on the map of Fig. 14.2, where it will be seen from north to south the several divisions are as follows; Lawrence uplift, Franks graben, Hunton anticline, Mill Creek syncline, and Tishomingo anticline. On Fig. 15.4 the Hunton anticline, Franks graben, and Lawrence uplift are combined under the general term, Hunton arch. The Arbuckle anticline is next south of the Tishomingo anticline but offset to the west. The Washita syncline and fault zone separate the Tishomingo anticline from the Arbuckle anticline. The structures are compressional in nature, and especially in the Arbuckle anticline and south-lying Ardmore basin thrust faults and tight folds are pictured by Dott (1934) and confirmed by Swesnik and Green (1950). The overriding is northward. Study sections in Fig. 15.6.

Ardmore Basin. The Ardmore basin is a folded and faulted basin between the Arbuckle anticline (Mountains) and Criner Hills. It contains a thick and deeply depressed Pennsylvanian sequence of clastic sediments, overlying a rather thick Cambro–Ordovician carbonate sequence with unconformable relations attesting two principal times of orogeny. These will be outlined presently.

Over 30,000 feet of Paleozoic sediments are involved, about 13,000 feet of which are Pennsylvanian and include the Springer, Dornick Hills, Deese, Hoxbar, and Pontotoc formations, from oldest to youngest. Most of the pre-Pennsylvanian beds are limestone, and the Pennsylvanian are mostly sandstone and shale. The Ardmore basin is considered a foredeep by van der Gracht, north of the Wichita and Criner Hills anticlinorium. At the time of deposition of the beds, the basin spread over the site of the present Arbuckle Mountains as well as the present Ardmore syncline, and extended to the Hunton–Tishomingo landmass (Dott, 1934).

Anadarko Basin. North of the Wichitas is the extensive Anadarko basin. It occupies the greater part of western Oklahoma. Its axis runs west-northwest and approximately parallels the Wichita–Amarillo uplift. The Permian beds thicken to 4500 feet just 25 miles north of the nearest granite outcrop. The thickness of the Pennsylvanian in the center of the basin is unknown but may be rather great, notably in the eastern part, and may be an extension of the Ardmore basin. Becker (1930) calculates the highest part of the Wichita anticlorinium to have been elevated structurally about 19,000 feet above the axis of the Anadarko "foredeep."

The Ardmore trough trends into the Anadarko basin under the blanket of Permian red-beds and Cretaceous. It is not known how much Pennsylvanian subsidence occurred in the Anadarko basin, but it is clear that most of the subsidence in the Ardmore basin is Pennsylvanian, and at least 4500 feet of subsidence in the Anadarko is Permian.

Paleogeology of the Wichita–Ouachita Region

The history of sedimentation in Oklahoma is in two distinct divisions, both in time and space. An uplift and peninsula through Texas from Mid-Ordovician to Mid-Mississippian separated the West Texas basin

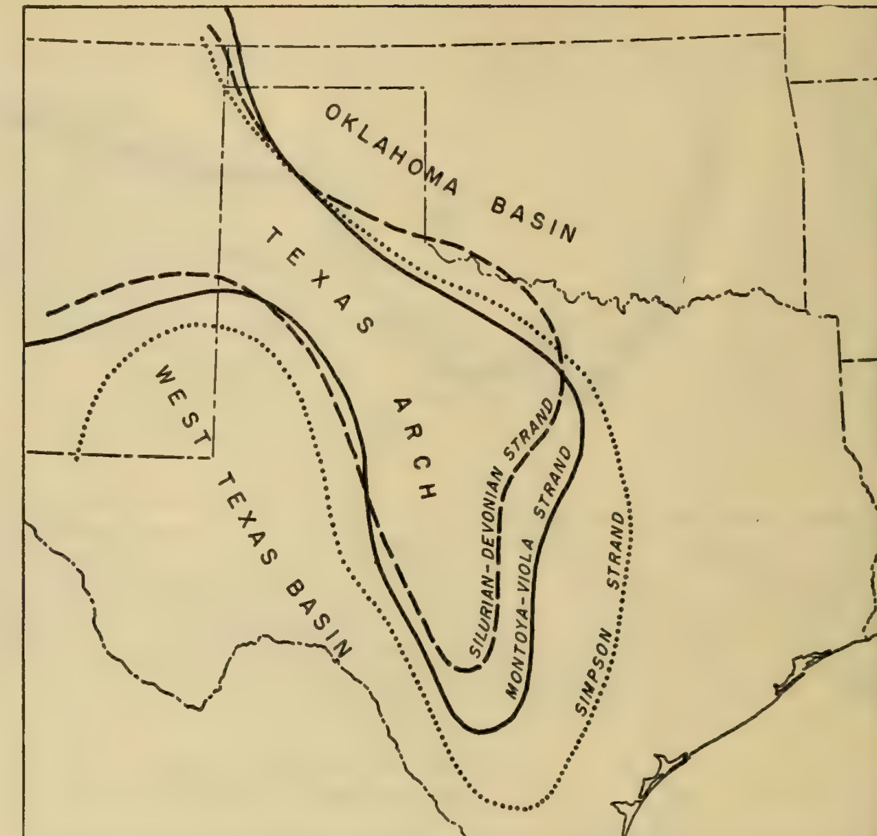


Fig. 15.7. Mid-Ordovician to Early Mississippian tectonic features of Texas and adjacent areas. After Adams, 1954.

from the Oklahoma basin. See Fig. 15.7. The Cambro–Ordovician Arbuckle limestone sea spread across the arch in platform fashion where about 1000 feet of carbonates accumulated, but in the Oklahoma basin on the northeast 4000 to 6000 feet accumulated. On top of these deposits, while the Texas arch was emergent, an additional 3000 feet of sediments were deposited in Late Ordovician, Silurian, and Early Devonian time. These were also mostly carbonates. To the north in Kansas the equivalent strata are only about 1000 feet thick. The region

of subsidence, defined by the pre-Mississippian strata, the Oklahoma basin, extended west-northwesterly toward the Colorado sag in central Colorado. The core area of the Ouachitas received about 3000 feet of sediments during this time, so the axis of the Oklahoma basin appears to have lain in the northern part of the Ouachitas and under the Arkansas Valley, and to have extended eastward to a connection with the Appalachian geosyncline. See Plates 2, 3, and 4.

In Late Mississippian time and during the Pennsylvanian a new regimen of sedimentation dominated the region, and over the carbonates and cherts great volumes of shales and sandstones were deposited. The basin of sharp subsidence and accumulation followed mainly the belt of later orogeny of the Ouachita system. The site of the present Ouachita Mountains, the Arkansas Valley and the Wort Worth basin marked the region of heavy deposition, but a spur of this arcuate basin projected off to the west through the Ardmore and Anadarko basins where at least 10,000 feet of clastic sediments accumulated. See Plate 8 of the Early Pennsylvanian. The history of Pennsylvanian sedimentation is complex because of deformational impulses from time to time and place to place. These will be discussed under the next heading.

Phases of Orogeny

Late Mississippian Phase. The great flood of Stanley, Jackfork, and Johns Valley clastics in the Ouachitas and the equivalent Springer group in the Ardmore basin reflect major uplift and associated deformation. This was a belt in the hinterland, toward the Gulf of Mexico, most probably, because there is no plausible source area to the north.

Early Pennsylvanian Phase. The first disturbance within the basin of accumulation is detected in a post-Springer and pre-Dornick Hills or Deese unconformity, in the Criner Hills–Ardmore basin area. See Fig. 15.6. This probably marked the beginning of rise of the entire Amarillo–Wichita element (Swesnik and Green, 1950). See tectonic correlation chart, Fig. 15.8.

The Ardmore basin then proceeded to sink and received an additional 17,000 feet of sediments making up the Dornik Hills, Deese, and Hoxbar groups.

Late Pennsylvanian Phase. By McAlester time (late Lampasas), the crest of the Hunton–Tishomingo uplift had been eroded to the Hunton limestone, while the northeastern flank was being submerged by encroaching seas. Erosion of the crest, due to intermittent uplifts, had exposed the Viola limestone. Then followed a rather extensive submergence, and by Missouri time the entire northwest end of the Hunton–Tishomingo landmass had been covered. The Ardmore basin, received sediments from the Hunton–Tishomingo and Wichita land areas as well as the previously elevated Ouachita Mountains. The basin spread over the site of the present Arbuckle Mountains. A series of rocks was deposited in this basin that differs in facies and sequence from the material that was being laid down simultaneously in the McAlester basin northeast of the Hunton–Tishomingo land mass. The two basins were probably never connected.

Late in Hoxbar time, compressive forces from the southwest rejuvenated the older folds of the Wichita uplift, and the entire element was moved northward toward the Hunton–Tishomingo buttress. The Amarillo Mountains were also rejuvenated, and the erosional detritus of the "granite wash," was formed. The sediments of the Ardmore–Arbuckle basin were greatly compressed, and the Arbuckle anticline originated. As the forces continued to move the southern elements northward, the eastern part of the Wichita system was thrust still farther north, apparently moving as a pivot, with the west end of the Amarillo Mountains remaining about stationary. The thrusting at the eastern end resulted in minor folds, first in the Arbuckle anticline and later in the Hunton–Tishomingo arch. Most of these minor folds became asymmetrical, and many were finally overturned toward the northeast. In the final stages of the movement, the major anticlines broke on their overturned axes, finally developing into thrusts and overriding the adjacent synclines.

Thirteen small erratic masses have been found toward the west extension of the Mill Creek syncline (Lehman, 1945). The erratics are remnants of an extensive thrust sheet which overrode at this place the truncated edges of the Simpson group in post-Hoxbar and pre-Pontotoc time.

In a detailed study of a small area in the Arbuckle anticline Dun-

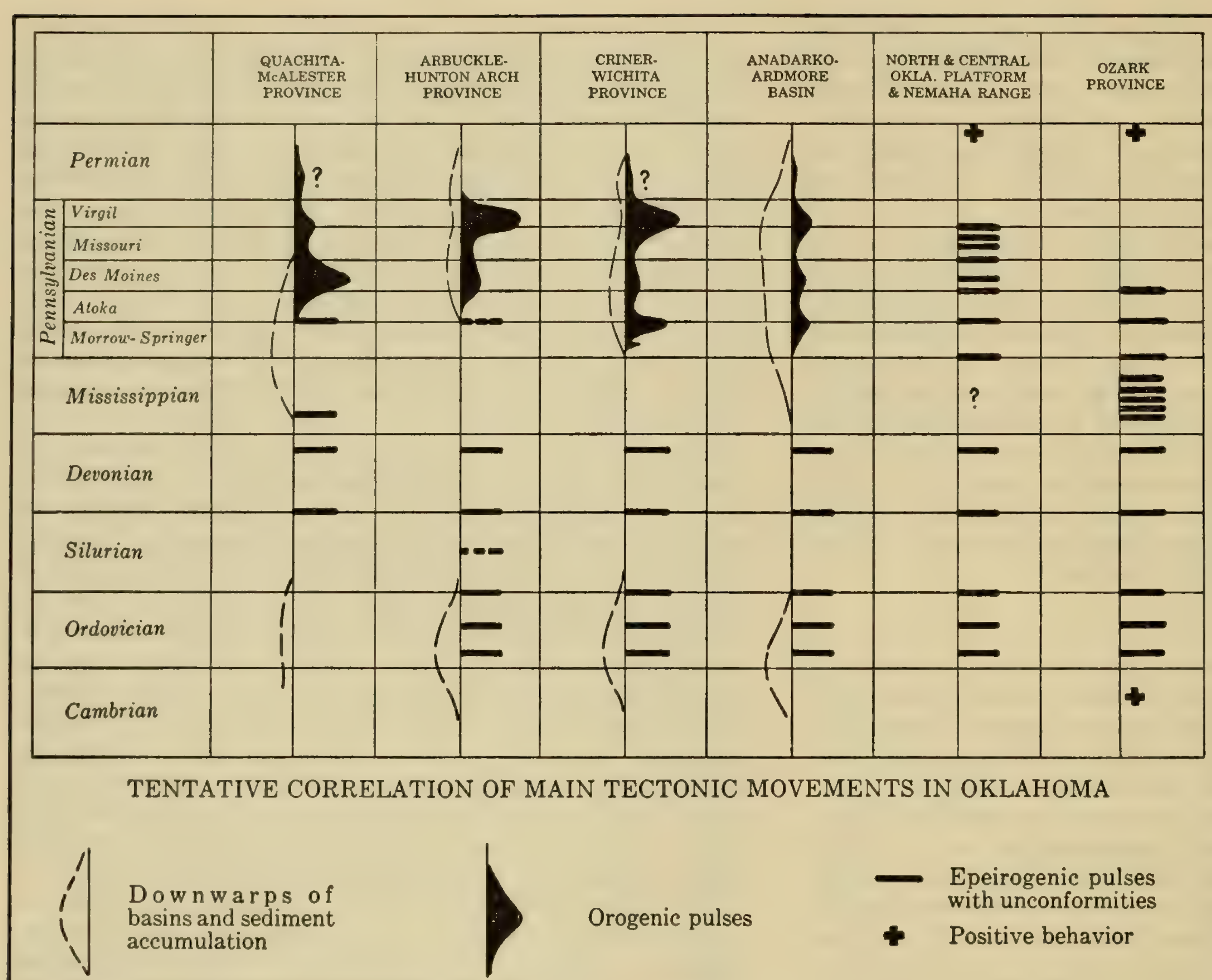


Fig. 15.8. Graphic representation of phases of deformation in Mid-Continent region. Reproduced from *Tectonic Map of Oklahoma*, 1956.

ham (1955) finds evidence by way of conglomerates, unconformities, and fault offset of fold axes that the deformation there began in Deese (Mid-Pennsylvanian) time, culminated in late Pennsylvanian, and continued on into early Permian by tilting the Lower Pontotoc conglomerate beds up to 40 degrees.

The Arbuckle anticline was thrust far northeast of its original position and overrode a considerable distance onto the Hunton–Tishomingo uplift. The magnitude of the overthrusting decreased a great deal within a short distance from southeast to northwest where crustal shortening was taken up mainly by complex folding. It probably follows that the thrust along its strike continued for a considerable distance southeast under what is now the Ouachita Mountains. The Tishomingo anticline was shoved northward in an overthrust second in magnitude only to the one in the southeast end of the Arbuckle anticline, and overrode the syncline to the north. The Franks graben, Wapanucka syncline, and other minor folds and thrusts were formed in the Hunton arch.

Tomlinson (1929) has estimated the amount of crustal shortening in the Ardmore basin as 16 miles; and Dott (1934), whose theory of structural evolution the above summary depicts, suggests in his illustration (Fig. 15.6) a net shortening at right angles to the trend of the structures, in late Pennsylvanian time only, of several scores of miles. It is, therefore, probably incorrect to show the positions of structural elements as they existed in times past in the places where the features now repose, but so many uncertainties attend the construction of palinspastic maps (Kay, 1945) in this region that it seems best for present purposes to crown the elements so as to conform to their present geographic positions. Such has been done on the tectonic maps of Plates 6, 7, and 8.

During the great late Pennsylvanian phase, marine deposits of Canyon age and older were highly tilted on the flanks of the Arbuckle Mountains, and during the following Cisco time erosion removed a sedimentary mantle probably 3 miles thick, and cut into granite. The granite thus removed was distributed in beds of Wolfcamp age over wide areas. North central Texas was affected to some extent at this time, as shown by thinning over the Matador arch.

TEXAS FORELAND

Definition

The Texas Foreland, as here defined is the fairly undeformed portion of the earth's crust north and west of the Ouachita–Marathon orogenic belt, south of the Wichita system, and west of the Laramide cordillera. See Fig. 14.1. It is characterized by broad arches, basins, platforms, and shelves. It appears to be a small part of the Central Stable Region cut off by the Wichita system. In reference to the Precambrian rock area Flawn (1959) has designated large parts of it as the Texas craton. Considerable igneous activity and probably deformation occurred in Precambrian time, but from the beginning of the Paleozoic era to the present it has been a fairly stable region with practically no igneous activity.

For purposes of discussion the Texas Foreland may be considered to have two divisions, the Central Texas and the West Texas–New Mexico.

Central Texas

Texas Arch. During Cambrian and Early Ordovician time Texas was mostly a shelf region of carbonate deposition. The carbonate deposit known as the El Paso limestone in New Mexico, the Ellenburger in West Texas and the Arbuckle in Oklahoma and adjacent north Texas, thickens southeasterly from a thin layer on the northwest to a massive deposit over 2000 feet thick at the edge of what may have been the continental shelf at the time. The Oklahoma basin lay to the north and the West Texas basin to the west. In about Mid-Ordovician time a broad and gently emergent peninsula extending southeastward through Texas rose (Adams, 1954). See the map of Fig. 15.7 and stratigraphic column of Fig. 15.9.

The subsurface outcrops as indicated on the map are interpreted to be depositional edges, with the peninsula, as large as Florida, emergent throughout the long period of time. The deposits in general gradually encroached on the peninsula; the Simpson, Viola, Montoya, Sylvan, and Hunter being Upper Ordovician and Silurian, and the Woodford Devonian. The lithologies are remarkably similar on either side of the arch.

STRATIGRAPHY

SYSTEM		SERIES	STAGE	GROUP	FORMATION	MEMBER
PENNSYLVANIAN SYSTEM	UPPER	CISCO		THRIFTY GRAHAM		
		CANYON		CADDO CREEK	Home Creek Ls Colony Creek Sh	
				BRAD	Ranger Ls Placid Sh Winchell	
				GRAFORD	Cedarton Sh Adams Branch Upper Brownwood Sh Palo Pinto	
				WHITT	Keechi Creek Salesville	Lake Pinto Ss
	MIDDLE	STRAWN			East Mountain Sh	Capps Ls
				LONE CAMP	Garner	Brazos River cong. Mingus Sh Thurber Coal
				MILLSAP LAKE	Grindstone Creek	Goen Ls Santo Ss Buck Creek Ss
					Lazy Bend (Restricted)	Brannon Bridge Ls Hill Creek
		LAMPASAS	KICKAPOO CREEK	"CADDO LIME" (Subsurface)	Rayville Parks Caddo Pool	Kickapoo Falls Ls Dickerson Sh
			ATOKA	"CADDO LIME" (Subsurface); BEND (Outcrop)	Smithwick	Lower "Caddo Ls" Lake Ss pay
				BEND (Outcrop)	Big Saline Upper Marble Falls	Brister Lemons Bluff Gibbons cong.
	LOWER	MORROW		BEND (Outcrop)	Comyn of Subsurface Lower Marble Falls	Aylor Sloan
		SPRINGER				

Fig. 15.9. Pennsylvanian stratigraphy of the Llano uplift. After Cheney and Goss, 1952.

Concho Arch. In late Mississippian time the orogeny of the hinterland of the Ouachitas resulted in the depression and fill of the Fort Worth and Kerr basins marginal to the later belt of compression. This resulted in the development of a broad, pronounced asymmetrical arch involving the previous formations and the top of the Precambrian. The situation is illustrated in the lower cross section of Fig. 15.10. Subsidence continued through the Atoka and Kickapoo clastics (subdivisions of the Lampasas according to Cheney and Goss (1952). The asymmetrical arch is called the Concho. With the deposition of the thick Permian sediments of the Midland basin (second cross section, Fig. 15.10) the arch becomes a very strong and large feature. The present contour of the Precambrian surface reflects the arch essentially as it was at the close of Permian time. See Fig. 15.1. It pitches gradually to the north-northwest and reaches to the Matador arch.

Bend Axis. The Permian and Upper Pennsylvanian beds overlap the Concho arch from the west in the manner illustrated in Fig. 15.11. As far as these beds are concerned an axis of down tilting to the west is involved, and this has been called the Bend axis or arch. There may be an arch in the Upper Pennsylvanian beds but probably not in the Permian.

Llano Uplift. The southeast end of the Concho arch was so high that all beds were stripped off down to the Precambrian before the invasion of the Cretaceous seas, which spread a cover of coastal plain sediments widely over the south and east flanks of the arch. These sediments have since been mostly removed from the Precambrian and a domal area known as the Llano uplift results. This is the most prominent feature evident on the geologic map of Central Texas.

The Pennsylvanian history of the site of the Llano uplift is somewhat more involved than the cross-sectional representation of the Concho arch in Fig. 15.10. According to Cheney and Goss (1952):

> Mississippian outcrops in the Llano region transgress the truncated Ordovician Ellenburger group. Drilling has shown an increasing loss of section west of the Llano uplift so that, as a result of both erosion and non-deposition, Upper Pennsylvanian (Canyon) marine sediments locally overlap Cambrian rocks in and near northeast Menard County. Farther west and northwest, Middle Pennsylvanian beds rest on truncated Mississippian and Ordovician or older rocks in a large region, heretofore called the Concho arch, where local as well as regional tectonic features had developed mainly along trends varying from north-northeast to northwest. Thin Middle Pennsylvanian marine sediments of the Lampasas and Strawn series deposited across this base-levelled region are chiefly limestones and shales of the platform type in contrast to thick basinal type deposits on the east and south.

A system of large faults extending northward from the present Llano uplift into the Fort Worth basin developed during very late Lampasas time. The faults, well known from surface mapping in the Llano uplift, have now been followed by geophysical work and drilling for more than 100 miles northward into the Forth Worth basin. Some of the faults have displaced upper Lampasas and older beds as much as 1100 feet in the

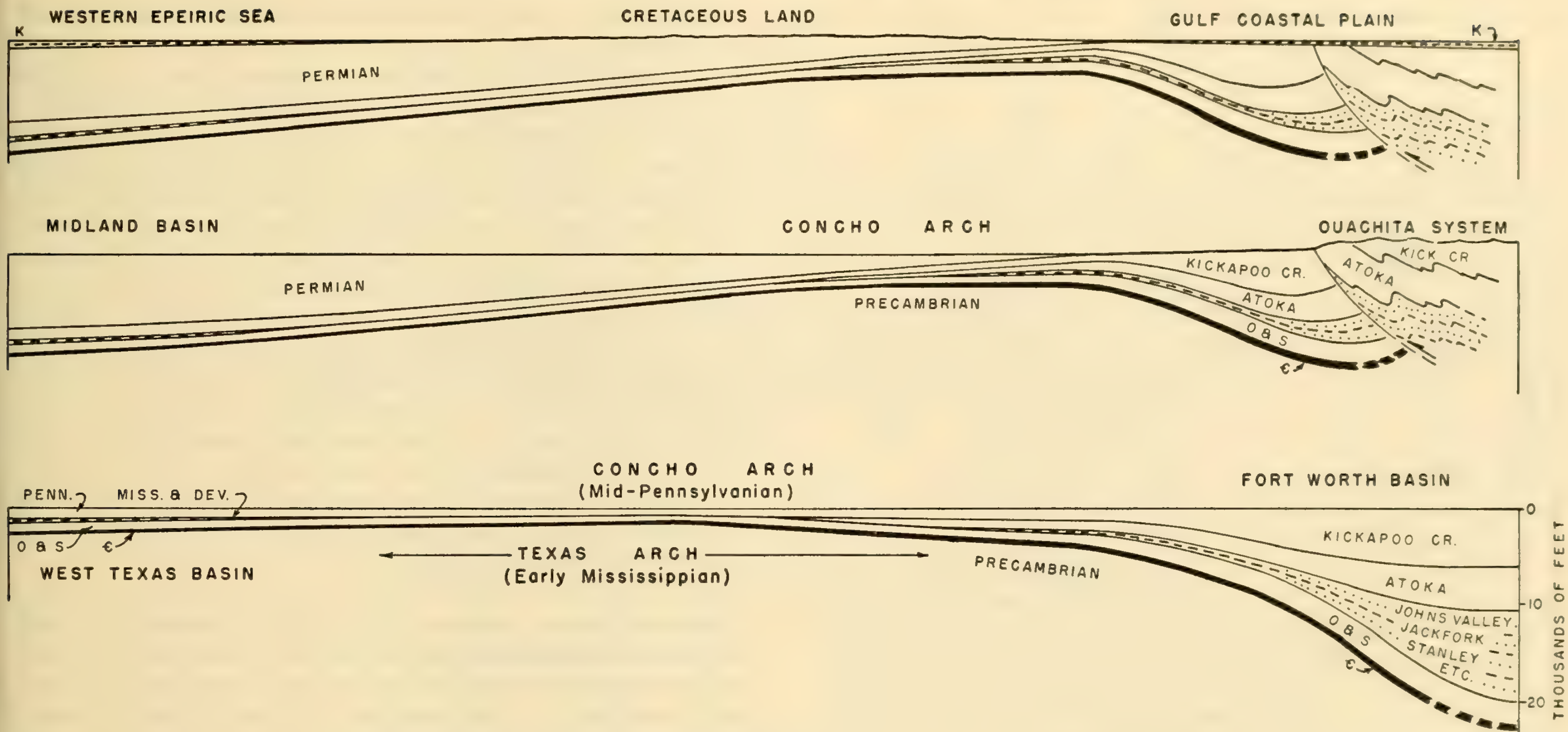


Fig. 15.10. Evolution of central Texas as idealized along an eastwest section from the Midland basin to the Ouachita orogenic belt.

Strawn basin. The faults in the Llano uplift have formed narrow grabens, and the three most prominent horsts over which the later strata are flexed are called, from west to east, the Richland Springs, San Saba, and Lampasas "axes." The Richland Springs axis forms the southern part of the present Bend arch. See map, Fig. 14.1.

The time of deformation of the Ouachita orogenic belt is believed to be post-Kickapoo. As cited in the treatment of the Ouachita Mountains a major unconformity across the Atoka, McAlester, Hartshorne, and Savanna beneath the Boggy shale in the west end of the Arkansas Valley is believed to mark the time of main deformation in the Ouachitas. This accords with Cheney and Goss's interpretation of the Pennsylvanian around the north and east sides of the Llano uplift.

West Texas–New Mexico Region

During Permian time, the foreland area in front of the Marathons was divided into a number of irregularly shaped provinces which received

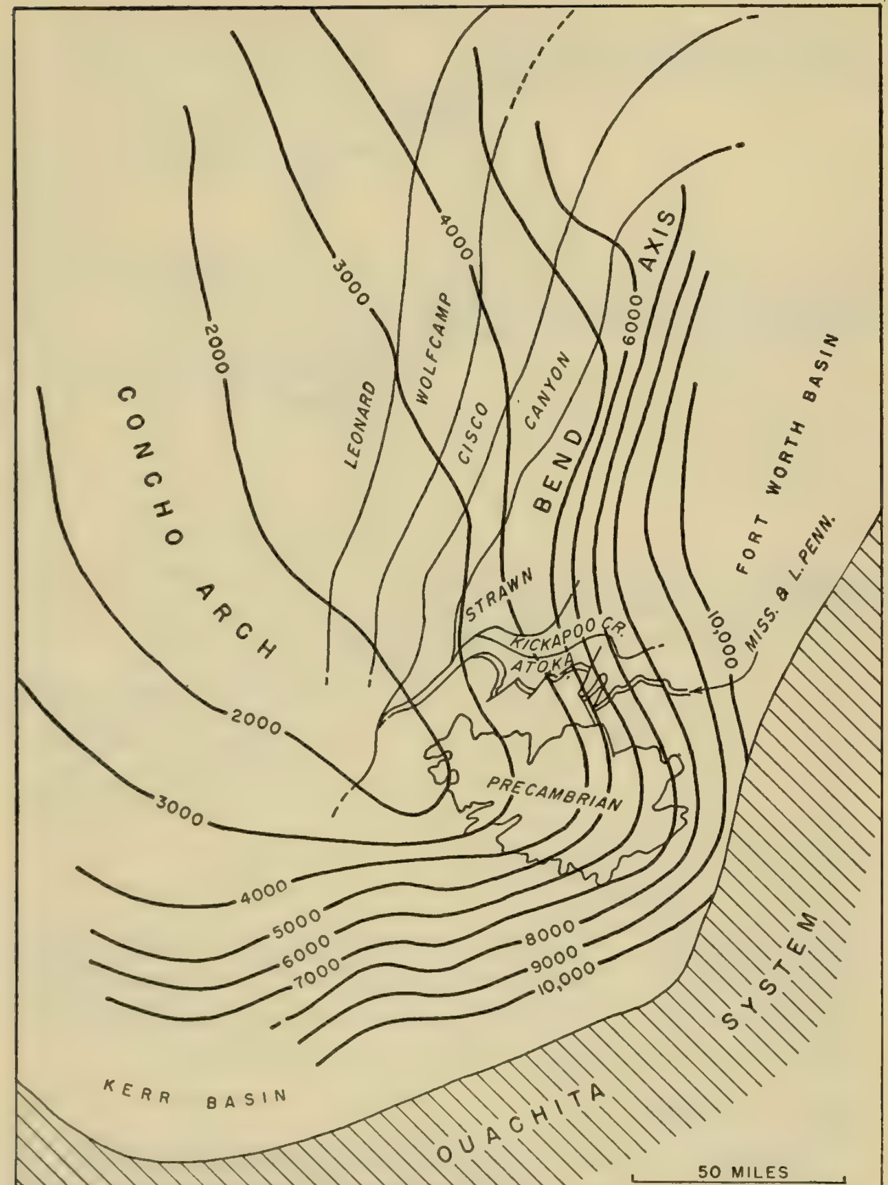


Fig. 15.11. Concho arch, Bend axis, and Llano uplift, after Cheney and Goss, 1952. Formational contacts generalized. Heavy contours are isopachs on the Paleozoic interval below the Strawn formation and illustrate the nature of the resulting Concho arch in Mid-Pennsylvanian time.

different types of deposits and which were probably tectonically unlike. Refer to Figs. 14.1 and 15.12. Some were basin areas, like the Delaware basin in which a total of 10,000 feet of sediments accumulated. Others were shelf areas. Akin to the shelves were several narrow masses, or platforms, lying between the basins. The basins were areas of greater subsidence; the platforms and shelves, areas of lesser subsidence. The Central Basin Platform was covered with 2000–4000 feet of sediments, as were also the shelf areas. The provinces appear to have been inherited from the pre-Wolfcamp foreland features, and each platform is underlain by one of the more important pre-Wolfcamp uplifts. The Permian tectonic features may have been formed during a time of dominant crustal tension, following the pre-Wolfcamp time of dominant crustal compression (King, 1937). The basins were centers of accumulations of clastic rocks, first black shales and later sandstones, and the total thickness of beds deposited in them was greater than elsewhere. Limestone tended to form over all the higher standing areas. Landward, because of climatic conditions that favored evaporation, evaporites were laid down in the fringing seas. On the margins of these seas, red-beds were deposited which were derived from the bordering lands.

The subsidence in Permian time that led to the burial of the Pennsylvanian ranges also resulted in the burial of the Matador and Amarillo–Wichita ranges to the north, and the northern part at least of the folded and thrust Marathons. Much of the sediment in the extensive Permian basin came from the Ouachitas which were actively being elevated at this time. Some debris from the Marathons reached surprising distances northward. The subsidence was regional in aspect and accentuated the Concho arch.

Extending across the larger features of the Marathon foreland and imparting a distinctive grain to their surfaces are numerous minor tectonic features in which the linear element dominates. These include the flexures in the Guadalupe Mountains region, the minor folds on which

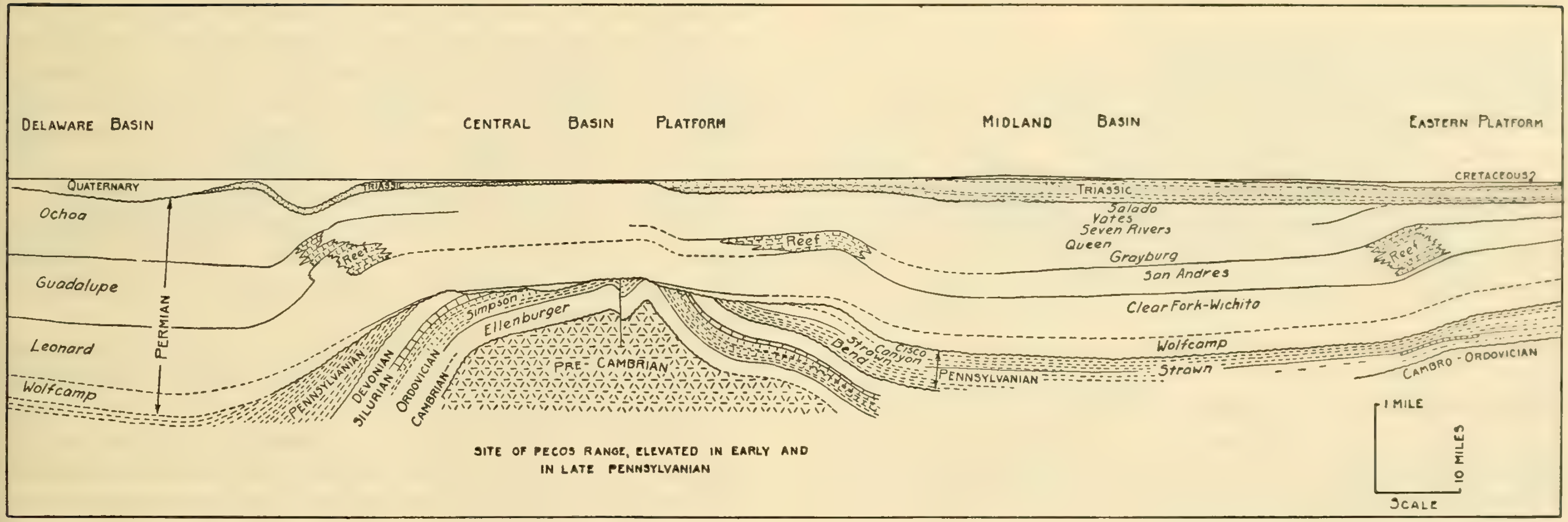


Fig. 15.12. Principal stratigraphic units and structural features of the South Permian basin of New Mexico and Texas. Line of cross section shown on map, Fig. 14.1. Taken from Plate 2, King *et al.*, 1942.

many of the oil fields are located, and various faults. Some were formed in pre-Wolfcamp time, but most of them were formed during the Permian. Some suffered movement again in the Cenozoic. They may be grouped into four systems, viz., northwest trending, northeast trending, north-northwest trending, and east-west trending; but much yet remains to be learned regarding the systems because those mentioned may not be natural units and may include some unrelated features. The systems apparently include features of several different ages, as well as features that were formed during several periods of movement.

ANCESTRAL ROCKIES SYSTEM

Major Structural Features

A group of imposing uplifts in Colorado and New Mexico of Pennsylvanian age, the Ancestral Rockies, has long been known. Considering the far greater length and breadth of the modern Cordillera known as the Rocky Mountains, the Ancestral Rockies only partially deserve their name. Deep basins are associated with the uplifts, and collectively represent a rather important orogenic system in the foreland. The Ancestral Rockies are separated from the Wichita system and the Texas foreland chiefly for purposes of discussion, but probably are continuous with and intimately related to them.

By reference to the map of Fig. 6.7 the several uplifts of the Ancestral Rockies and the adjacent basins may be seen. Two of these, the Uncompahgre and Colorado, were particularly bold and high. The Colorado Range is frequently referred to as the Front Range highland. The Pedernal uplift is not yet very well defined, but seems to be an emergent area in east-central New Mexico which connects southward with the Diablo uplift. The Zuni uplift, like the Pedernal, seems to have been wide and not particularly high.

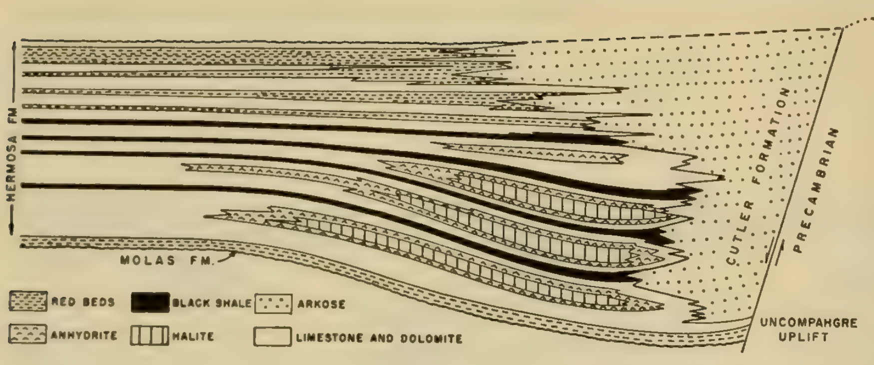


Fig. 15.13. Pennsylvanian deposits of the Paradox basin. After Herman and Barkell, 1957.

Pre-Pennsylvanian Setting

The total thickness of the Paleozoic formations present in central Colorado by the end of Lower Mississippian (Leadville) time was only 1000 feet. In southwestern Colorado, only 400–500 feet existed, and in the northern part of the Front Range, they were still thinner. Since the fairly pure Mississippian limestones occur in areas close to the Pennsylvanian ranges and no lithologic changes are evident in the limestones as the ranges are approached, the Mississippian seas probably spread over the sites of the highlands (Lovering, 1933).

Evidence of thinning, probably by erosion, is evident when isopachs are worked out, and it appears that some of the ranges first began to be expressed in latest Mississippian time, as illustrated in Fig. 6.6. The New Mexico arch of Mississippian age exposed Precambrian rock over much of central New Mexico.

Uncompahgre and Colorado Ranges

The Uncompahgre and Colorado ranges were flanked by basins as indicated on the map of Fig. 6.7; the Paradox, the Central Colorado, and the Denver. The extreme and abrupt facies changes of sediments deposited against their flanks is the evidence of the sharp uplifts. One of the flanking basins, the Paradox, is illustrated in Fig. 15.13. The southwest margin of the Uncompahgre Range was a fault scarp, and the thin pre-Pennsylvanian sedimentary veneer was soon stripped from the rising block, with the Precambrian crystallines furnishing flood deposits of arkose to the adjacent subsiding basin. During part of Pennsylvanian time evaporite conditions prevailed and four evaporite sequences—cyclothems—resulted (Herman and Barkell, 1957). This part of the Hermosa formation is the Paradox facies or member.

The Molas is Atoka in age and the Hermosa spans the Des Moines, Missouri, and Virgil. The Cutler extends on into the Permian. The time of the most vigorous uplift, then, is clearly Atokan through to the beginning of the Permian.

The Pennsylvanian and Permian sediments overlap the gently beveled edges of the older Paleozoic rocks and rest on Precambrian crystallines in the cores of the old ranges (Lovering, 1933; Burbank, 1933; Glockzin and Roy, 1945). See Fig. 15.14. The crystalline rock was the source of many of the Pennsylvanian and Permian strata which are commonly coarse and arkosic near the old landmasses. For instance, Brill (1944) describes the sediments of the central Colorado basin as consisting mostly of red and gray arkoses, arkosic conglomerates, sandstones, siltstones, and gypsum which thicken to 13,000 feet in the deepest part of the basin. Lateral variations are abrupt and extreme. During the most active time of uplift of the adjacent ranges, the coarse clastics were deposited as deltas along the margins of the trough, and the fine-grained sediments were carried into the center. Identical mineral assemblages in the clastics on both sides of the basin indicate that the exposed bedrock of both the Uncompahgre and the Colorado Range was much the same.

Pedernal Uplift

The Pedernal landmass, named by Thompson (1942) from the Pedernal Hills, is a large north-south-trending range in east central New Mexico, about midway between the Rio Grande and the Pecos rivers. Red shales, sandstones, variegated shales, and limestones of Permian age rest directly on igneous and metamorphic rocks of Precambrian age in an area extending from the eastern side of the Sacramento Mountains, Otero

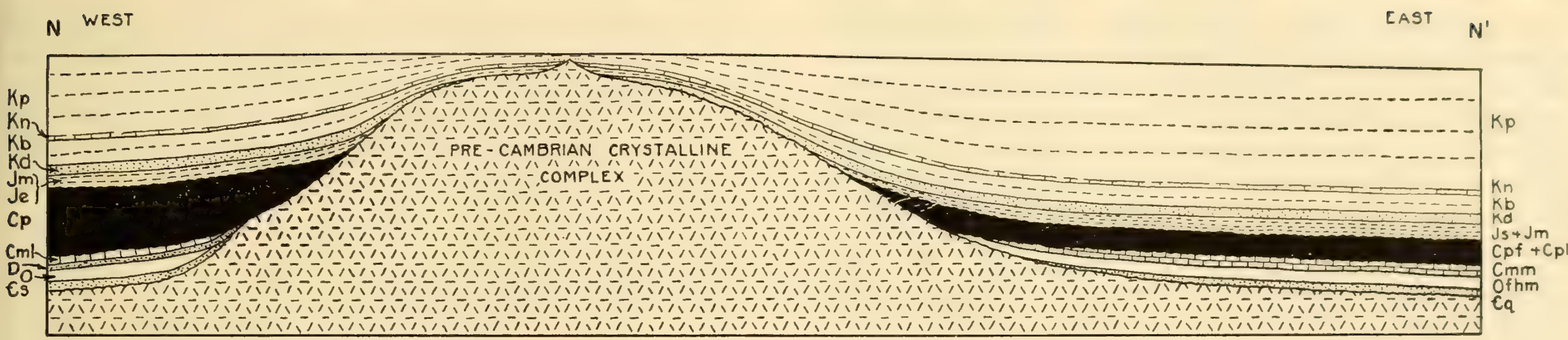


Fig. 15.14. Idealized section after Lovering (1929) of the Colorado Range near the close of Pierre time (Upper Cretaceous) and before deformation during the Laramide revolution. Kp, Pierre sh.; Kn, Niobrara ls.; Kb, Benton sh.; Kd, Dakota ss.; Jm, Morrison fm.; Je, Entrada ss.; Js, Sundance fm.; Cp, Belden sh.; Maroon fm. (Des Moines) and State Bridge siltstone (Permian); Cml, Leadville ls. (Mississippian ?); DO, Devonian and Ordovician formations; Ꞓs, Sawatch quartzite; Cpf and Cpl, Fountain fm. and Lykins fm. (Pennsylvanian and Permian ? red arkosic ss. and congl.); Cmm, Millsap fm. (Mississippian); Ofhm, Fremont ls. (Ordovician); Ꞓq, Cambrian quartzite. Vertical scale exaggerated and relative thicknesses of formations only approximate.

County, apparently continuously to northern Torrance County. Very coarse conglomerates with cobbles of quartzite and other metamorphic rocks are present in the Pennsylvanian rocks in the Sacramento Mountains on the west side of the uplift. From these and other similar data, Thompson concludes that the Pedernal Range was in existence from early Pennsylvanian time until well after the beginning of Permian time.

The Colorado Range probably extended southward into New Mexico through Colfax and Mora counties. Very thick sections of Pennsylvanian rocks of Des Moines age or older crop out on the eastern edge of the Sangre de Cristo Range from the region of Pecos River almost to the Colorado border. These rocks include arkoses, arkosic conglomerates and sandstones, and black shales.

Zuni Uplift

In northwestern New Mexico and northeastern Arizona evidence of Pennsylvanian uplift is noted in the modern Zuni and Defiance ranges. With additional subsurface evidence from drilling the configuration of the range appears to be like that illustrated in Fig. 6.7. Red sandstones and shales identified as the Permian Ajo formation rest on Precambrian crystalline rock in the Zuni Range, and the Permian Moenkopi formation rests on the crystallines in the Defiance Range. The old uplift is designated both as the Zuni and Defiance, but Zuni seems to be preferred.

Florida Uplift

In the Florida Mountains of southwestern New Mexico, Permian limestone rests on Ordovician limestone; and a short distance to the north in the Cooks Range, the Permian rests on Mississippian limestone. The absence of Pennsylvanian strata is due to a Pennsylvanian or post-Pennsylvanian disturbance, probably at the same time as those in the south end of the Hueco Mountains and the Diablo Mountains near El Paso. The direction in which the elevated land trends is believed to be southeasterly. It will be called the Florida Range.

Burial of the Ancestral Rockies

During Triassic, Jurassic, and Cretaceous time the Ancestral Rockies were gradually buried by accumulating sediments. Immediately around them were their own waste products, but marine epeiric seas brought carbonates and fine clastics from adjacent regions, and these sediments

helped in the burial process. Jurassic desert conditions brought great volumes of wind-transported sand from the western Cordilleran geanticline.

The Zuni and Pedernal uplifts were buried by the Permian deposits, the Uncompahgre lasted in a number of small islands until late Jurassic before final burial, and the Colorado Range lasted until Pierre time in the early Upper Cretaceous before the last peaks were drowned. See Fig. 15.14. By this time the early manifestations of crustal unrest in the Rocky Mountains are evident, and the old buried ranges with overlying sediments were considerably deformed. An array of new superposed structures and ranges developed, which will be reviewed in later chapters.

16.

THE LATE PALEOZOIC ZONES OF FAULTING AND CRYPTOVOLCANIC OR METEORITE IMPACT STRUCTURES

FORELAND ARCUATE FAULT ZONE

An arcuate zone of faults extends from the Llano dome in Texas northward to Oklahoma, northeastward to the Ozark dome and eastward across Kentucky to West Virginia. The faults are subparallel with the zone for the most part, but some are divergent, especially two long faults in Missouri which strike northwestward directly athwart the zone. See Tectonic Map of the United States, 1944. The zone crosses domes and basins alike, and therefore does not seem to be controlled by them. On the other hand, the fault zone wraps around the Ouachita arc of the marginal orogenic belt of the continent, and although the fault zone has a lesser curvature than the Ouachita arc and departs from it a considerable distance on the north, the subparallelism may mean that a genetic relationship exists.

Most of the faults are known to have originated in Pennsylvanian time or immediately thereafter. A few others are post-Devonian. Thus the time relation as well as the spatial indicates that the zone of faults is a single tectonic element.

The faults in the Llano uplift are known from surface mapping. Oil wells and geophysical prospecting have extended the known length of some of the faults more than 100 miles to the north-northeast into the Strawn basin (Cheney, 1940). The faults are probably of the high-angle, normal variety, and have blocked out narrow grabens and horsts. The high blocks have been named from west to east, the Richland Springs, Pontotoc, San Saba, and Lampasas axes. According to geophysical work in the Fort Worth basin, some of these faults have displaced the Smithwick formation 1100 feet, so the movement occurred in post-Smithwick (Early Pennsylvanian) time. However, beds only slightly younger than Smithwick, namely middle Strawn, are only slightly disturbed along a major fault near Regency in the Colorado River area, and the faults are not known in still younger Pennsylvanian beds. Cheney, therefore, concludes that the faulting in the Llano dome and Strawn basin occurred in early Pennsylvanian time.

The Stonewall fault in the Hunton arch area of southern Oklahoma is said to have occurred in about middle Strawn time and to have a displacement of about 3500 feet (Morgan, 1924), but from Dott's (1934) discussion the fault may be one of the Arbuckle group of thrust faults and not a part of the arcuate fault zone.

A group of faults in the northeastern corner of Oklahoma bound six small crustal blocks, each about 6 miles wide (Wilson, 1937). The faults trend in general northeastward, and some have been traced for 50 miles. One block is tilted to the north; two are tilted to the south, and the remaining three are about level. Their throws range from 90 to 600 feet, but these figures apply only to surface offsets. The faults, as well as the folds in this region, become more pronounced in the older underlying formations. Where it is possible to trace the stratigraphy at depth by oil wells, the structural relief decreases upward through the conformable sand-

stones and shales of the Atoka, Hartshorne, McAlester, Savanna, and Boggy formations. Bloesch (1919) believed that this decrease is due to recurring deformation during the deposition of the above formations, which are of the late Early Pennsylvanian age.

North of the six fault blocks and parallel with them is the hundred-mile-long Seneca fault. It extends into the southwest corner of Missouri. Surface evidence for the fault is not conclusive everywhere along the structure, and it may be a syncline rather than a fault in several places (Weidman, 1939).

Several rows of small faults are well known in Creek and Osage counties, Oklahoma, just west of the previously mentioned Seneca fault. The individual faults are arranged *en echelon* and trend northwest. The rows trend nearly north and make an angle of about 45 degrees with the individual faults. On Plate 8 the rows are indicated by dashed lines. The largest stratigraphic throw at the surface is about 130 feet, and the greatest length is 3¼ miles; yet the length of one of the rows is 80 miles. They are all normal faults.

Fath (1920) analyzed the *en echelon* faults as follows. The Precambrian crystalline rocks were cut by a system of faults before the Paleozoic veneer of sediments was laid down. In fact, peneplanation had removed most of the relief incident to the faulting before the Paleozoic beds started to accumulate. Beginning in Early Pennsylvanian time, the faults in the basement complex again became active, this time, however, with horizontal (strike-slip) movement. The overlying veneer was shear-strained along the underlying faults and broke in rows of small faults arranged *en echelon*. The east side of each fault in the Precambrian moved northward. The movement recurred several times during the Pennsylvanian, so that the throw of the faults is greater at depth. Some rows of *en echelon* faults may not even show in the Pennsylvanian beds at the surface today, and others are reflected in small asymmetrical anticlinal flexures over the faults. Several such rows of anticlines farther west in Oklahoma and north in Kansas may belong to the same system.

The postulate that the great arcuate fault zone is related to the Ouachita lobe of the marginal orogenic belt is supported by Fath's mechanical analysis. As would be expected in this theory, the foreland block directly in front of the lobe would be moved horizontally ahead of its left-hand neighbor, which is the direction of shear indicated by the *en echelon* faults.

In eastern Missouri, two stages of faulting are recognized (Weller and St. Clair, 1928), one in late Devonian and one in post-Mississippian. The faults form a complex system, and the total displacement in the fault zone ranges up to 1200 feet.

The eastern Missouri faults continue eastward across the southern tip of Illinois into Kentucky, where a region of widespread and intensive faulting exists. Along the north side of this complex of faults is the Shawneetown fault of southern Illinois, and its eastward continuation, the Rough Creek fault zone of Kentucky. The Shawneetown–Rough Creek fault zone is really an uplift that varies in structural relief and detail from place to place (McFarlan, 1943). Most characteristic of the uplift is its anticlinal structure. At places, a series of anticlines is developed, in part asymmetrical to the north and broken to form reverse faults. Normal faults arranged *en echelon* are also present. The structural relief of the uplift ranges from a few hundred feet to 2500 feet, and Mississippian beds in places are brought into outcrop. The complex structural zone divides the Pennsylvanian coal basin into a northern and a southern division.

Just south of the Shawneetown–Rough Creek structure in western Kentucky is a cluster of high-angle faults, the main ones of which trend northeast and east and have displacements up to 1500 feet. They are joined by smaller cross faults. The area is semicircular, about 60 miles in diameter, and is the most intensely faulted area in the interior lowlands. Along with the faults, peridotite dikes and highly commercial veins of fluorspar occur. The faults are post-Pennsylvanian and pre-Cretaceous.

The Rough Creek fault zone is continued after a gap of a few miles to central and eastern Kentucky by the Kentucky River fault and its associates, the West Hickman fault, the Irvine–Point Creek fault, and other smaller ones (McFarlan, 1943). Maximum displacement on the Kentucky River fault is 600 feet. Some suggestion of pre-Pennsylvanian movement and progressive movement has been made, but McFarlan believes the faulting occurred in post-Pennsylvanian time.

LAKE SUPERIOR FAULT ZONE

The *Tectonic Map of the United States* shows a group of long and subparallel faults extending from the Lake Superior region southwestward into Wisconsin and Minnesota. The Keweenawan fault is probably the best known. It runs lengthwise and approximately in the center of the Keweenawan peninsula of Michigan and separates the copper-bearing Keweenawan volcanic series from the Cambrian (?) sandstones. The fault is clearly a thrust in one exposure near Houghton, but probably a fairly high-angle one, with the Keweenawan series displaced southwestward over the Cambrian (?) sandstone.

North of the Keweenawan fault, the volcanic series is downfolded into a broad syncline with dips on the southeast flank of about 30 degrees. The beds rise and crop out on Isle Royal in Lake Superior. A fault which cuts the north flank of the syncline has been postulated just north of Isle Royal. This northern fault has been connected with the Douglas fault, which runs almost east-west south of Superior, Wisconsin, and which, according to Thwaites (1912, 1935) dips 38 to 60 degrees southward. He believes the south block has been thrust northward 6 to 12 miles. The Douglas fault swings southward after entering Minnesota, and there it has been studied by geophysical means. Near Pine City, the fault is believed to be nearly vertical, with the east side upthrown about 9000 feet (Welch, 1941).

The great syncline between the oppositely dipping Keweenawan and Douglas faults is thought by Thwaites to contain numerous minor folds in Wisconsin and hence to be a synclinorium. The structure is discussed in Chapter 4 and illustrated in Figs. 4.3 and 4.7. He also states that part of the displacement could have occurred in late Keweenawan time, but that part of it probably occurred later. The complementary relation of the Keweenawan and Douglas faults suggests they are of the same age. The syncline in the Keweenawan peninsula region appears to have subsided partly at the time the volcanic flows and conglomerates were accumulating, according to Broderick (personal communication), but considerable faulting undoubtedly occurred later. Thwaites (1943) agrees in substance with this view.

A fault along the north coast of Lake Superior has been surmised, chiefly on physiographic evidence (Martin, 1908), but this is not supported by gravity surveys.

Ten miles southeast of the Keweenawan fault in Michigan, two hills, Limestone Mountain and Sherman Hill, are made up of a basal sandstone and overlying dolomites and limestones. The sandstone, according to Case and Robinson (1915) is either Cambrian or Lower Ordovician, and the limestones and dolomites are Ordovician, Silurian, and Devonian. According to Thwaites (1943) the sandstone is Upper Keweenawan, and the dolomites and limestones are Trenton–Black River. The strata are cut by small faults and, along the east side, exhibit dips up to 55 degrees. A major fault may exist along the east margin, and the high dips may be drag along the fault which would be approximately parallel with the Keweenawan. The Ordovician beds in Limestone Mountain are 80 miles from the nearest Ordovician outcrops; and the Devonian, if present, 150 miles from the nearest Devonian outcrops.

Dating the faults in Limestone Mountain and Sherman Hill is difficult because of lack of agreement on the age of the sandstones associated with them (Cambrian or Precambrian), and the extensive swamp and drift cover that prevents working out the geologic relations. Opinion seems to favor an early episode of subsidence in which the Keweenawan basins were formed, and a later episode of faulting in which the Paleozoic rocks were affected.

The disposition of an immense amount of material that came from the truncation of thousands of feet of Keweenawan strata along the Douglas and Keweenawan faults poses another problem. If most of the movement were Precambrian, representative deposits possibly should occur in the Cambrian, but the orogenic waste products do not seem to make up any of the Paleozoic rocks nearby in Wisconsin or Michigan.

If all but a small part of the faulting were Precambrian and associated with the downfaulting of a basin in which the Keweenawan series accumulated, and if the Keweenawan series is 1100 m.y. old as recounted in Chapter 4, then during the next 500 m.y. before the basal Cambrian sands were spread across the region, all relief could have disappeared,

and the Cambrian sediments need not necessarily contain the Keweenawan lithologies.

CRYPTOVOLCANIC OR METEORITE IMPACT STRUCTURES

Definition

Eight small circular structures, one of known volcanic origin, and the others supposedly of volcanic origin, have been mapped in the Central Stable Region of the United States, and possibly a ninth one in the Colorado Plateau of Utah. See map of Fig. 16.1. They are described by Bucher (1933) as characterized by a nearly circular outline, a central uplift with intense structural derangement, and a marginal, ring-shaped

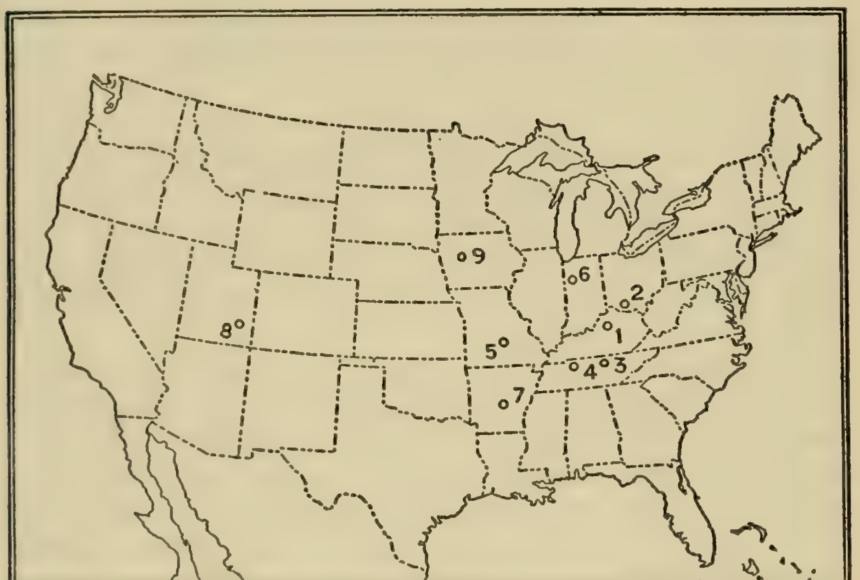


Fig. 16.1. Cryptovolcanic or meteorite impact structures in the United States. 1, Jeptha Knob, Shelby County, Ky.; 2, Serpent Mount structure, Adams and Highland counties, O.; 3, Flynn Creek disturbance, Tenn.; 4, Wells Creek basin, Houston and Stewart counties, Tenn.; 5, Decaturville structure, Camden and Maclede counties, Mo.; 6, Kentland structure, Newton County, Ind.; 7, Magnet Cove, Hot Springs County, Ark.; 8, Upheaval dome, San Juan County, Ut. After Bucher, 1933. 9, Manson, Ia.

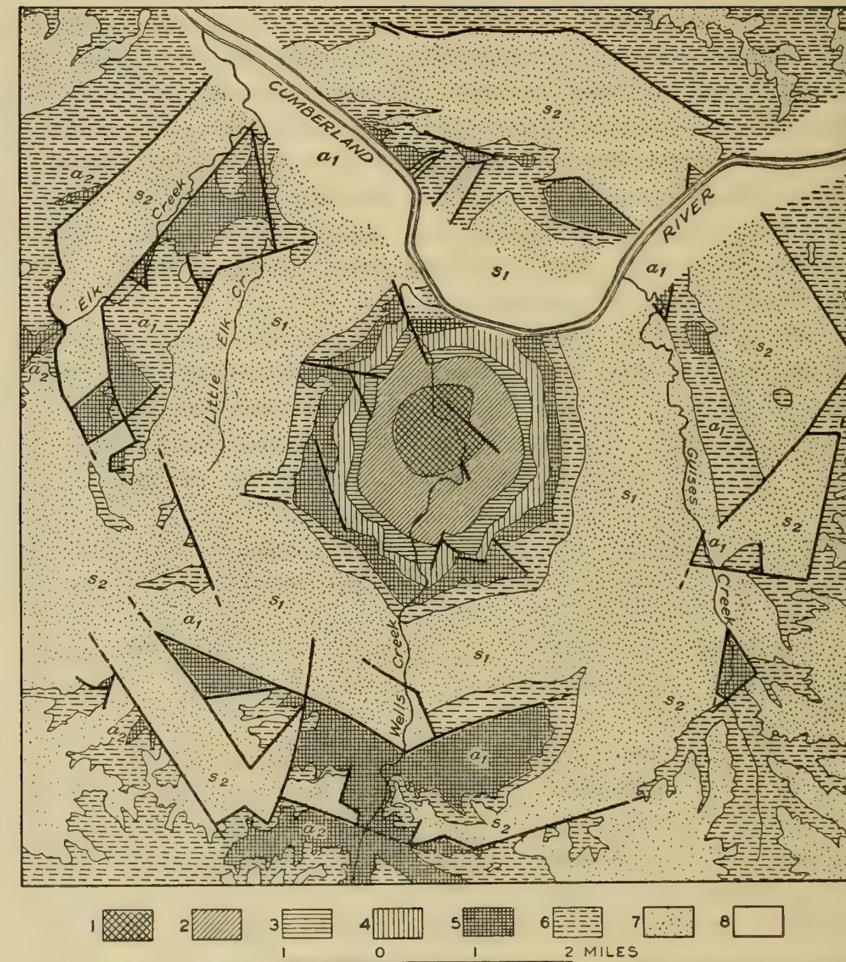


Fig. 16.2. Geologic map of the Wells Creek basin, Tenn. Reproduced from Bucher, 1933. 1, Wells limestone (L. Ordovician); 2, mid-Ordovician limestone; 3, Hermitage formation (mid-Ordovician); 4, Silurian and Devonian formations; 5, Lower Mississippian formations; 6, Warsaw limestone (mid-Mississippian); 7, St. Louis limestone (mid-Mississippian); 8, alluvium.

depression with irregular and local faulting. Including the marginal ring, they range in diameter from 2 to 8 miles. The inner intensely deranged core may be only part of a mile across in some, but in others up to 2 miles.

The faults make both an approximate concentric pattern and a radial pattern. In some, the radial pattern is resolved strongly into a northwest-southeast direction. Examine the representative illustrations of Figs. 16.2 and 16.3.

Distribution

The map of Fig. 16.1 shows the distribution of the cryptovolcanic structures. Numbers 1 to 6, and 9 are in the general arch and dome area of the central Mississippi Valley. Numbers 1 to 4 are in the Cincinnati and Nashville domes, number 5 in the Ozark dome, and number 6 in the Kankakee arch. They avoid the Illinois basin fairly well. Number 7 is in the orogenic belt of the Ouachita Mountains, but it is dissimilar to the rest in having igneous rocks exposed in the core, in being free of faults, and in being the site of considerable mineralization. See Fig. 14.2. Number 8 is in the Colorado Plateau and is complexly associated with salt dome upheaval.

Origin and Age

No volcanic rocks are associated, at least at the surface, with the small circular structures of the Central Stable Region, yet their circular shape, their upheaved, broken, and in places brecciated condition, and the presence of a number of dikes cutting the near horizontal Paleozoic rocks in surrounding areas, lead Bucher to imagine an explosive volcanic origin.

> These cryptovolcanic structures are thought to be the result of a sudden liberation of pent-up volcanic gases, which had accumulated near the surface, the explosion having been too weak to produce a shallow crater such as formed in the Ries Basin, southern Germany (Bucher, 1933).

These unique structures in the United States have been eroded more than those of Tertiary age in Germany, and so Bucher regards them as older and of probable late Paleozoic or Mesozoic age.

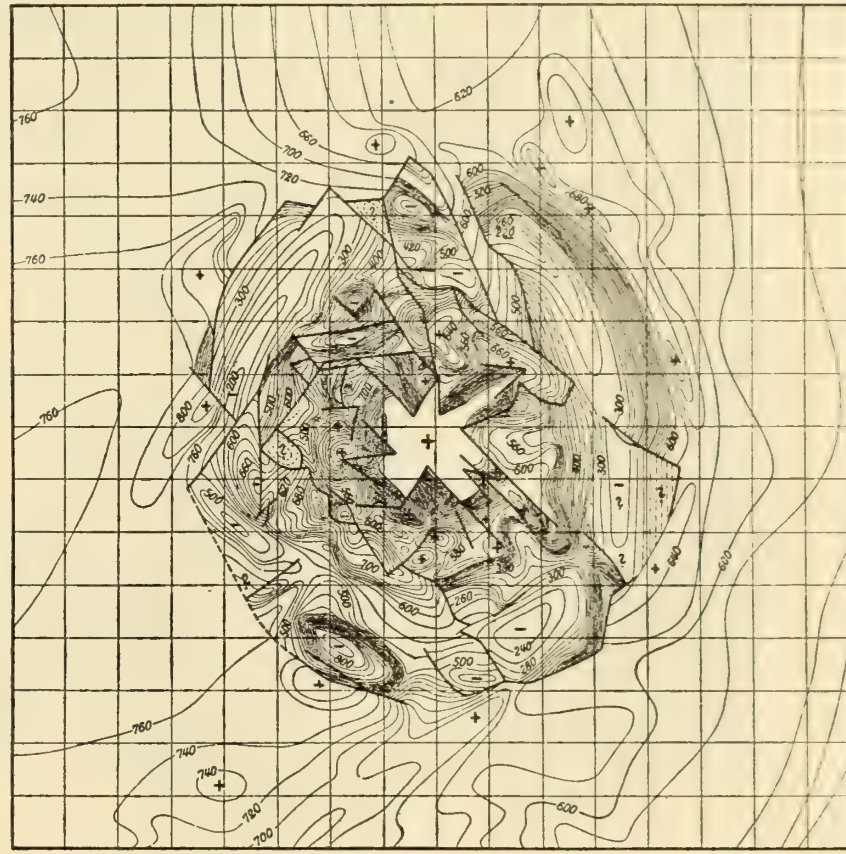


Fig. 16.3. Structure contour map of Serpent Mound, O. The length of each square is about 2200 feet. Reproduced from Bucher, 1933.

Recently a new cryptovolcanic (?) structure has been found near Manson, Iowa. It is number 9 on Fig. 16.1. Unlike the others it has a Precambrian crystalline core about 1½ square miles in area which lay unknown because of a cover of glacial drift until discovered by core drilling (Hoppin and Dryden, 1958). In this area a thin Paleozoic veneer of sedimentary rocks plus a cover of Cretaceous shale is the normal expectation under the drift. The contact of the crystalline rock with the surrounding sedimentary rocks dips outward 350 feet per mile to the

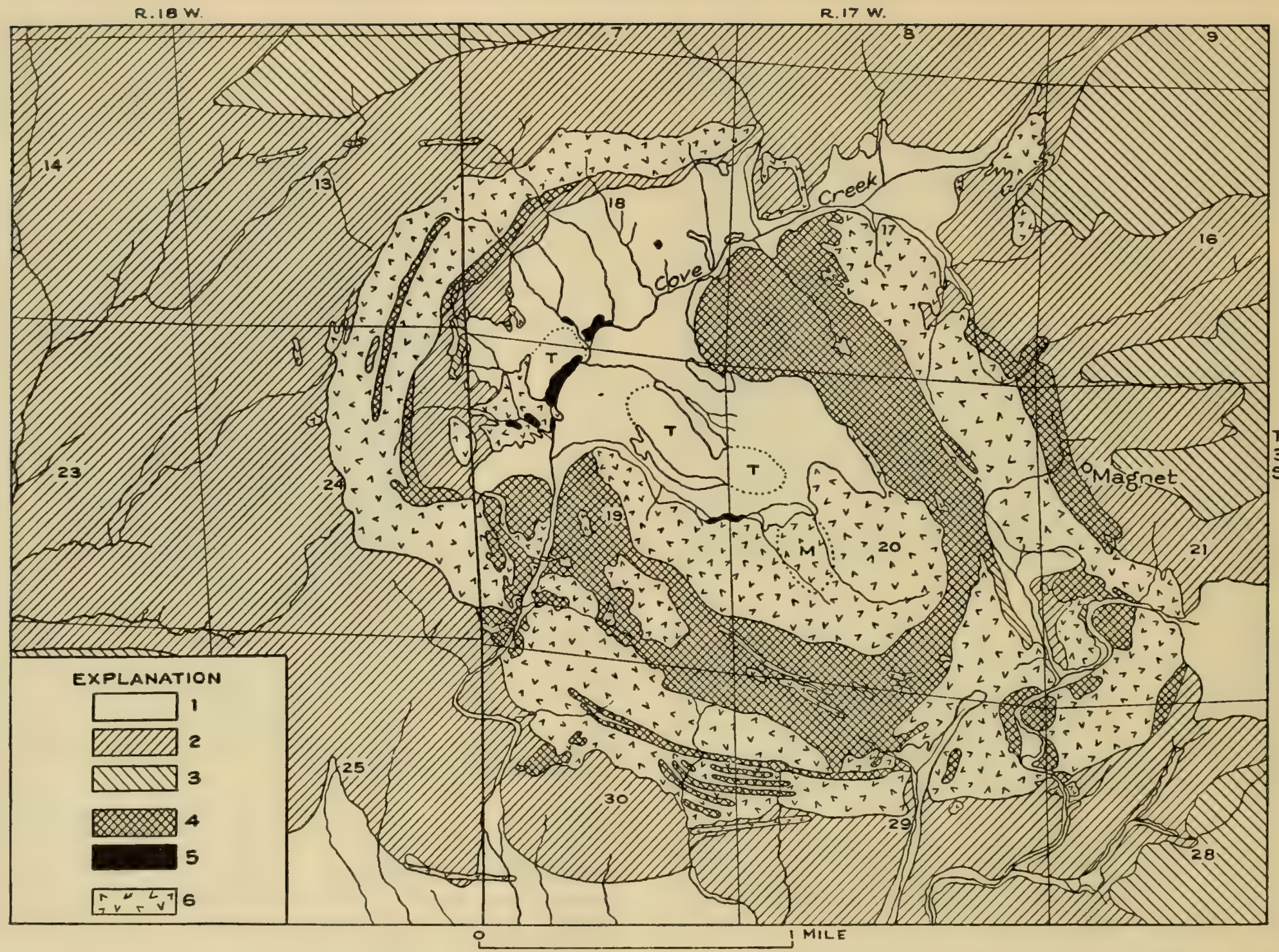


Fig. 16.4. Geologic map of Magnet Cove, Arkansas. Reproduced from Bucher, 1933; after Landes, 1931. 1, Pleistocene (T, tufa); 2, sandstone and shale (Mississippian); 3, novaculite (Devonian); 4, metamorphosed sandstone and shale; 5, metamorphosed limestone; 6, igneous rocks (M, magnetite).

southeast and 290 feet per mile to the west. The relief of this rather sharp, small dome is at least 1500 feet. Surrounding the crystalline core is a "disturbed area" 20 miles in diameter in which the sedimentary rocks are severely deformed. Mississippian limestone and Lower (?) Cretaceous shale have been sampled in drill cores in the disturbed area, and therefore the structure was formed in post-Early Cretaceous time. The upper 200 feet of the Precambrian rock beneath the drift is badly shattered.

The writers believe that the crystalline rock was forced upward into the limestone and shale, and in the process was badly shattered. The mechanism responsible is postulated to be a hidden igneous intrusion.

Magnet Cove

The Magnet Cove structure is included by Bucher in his résumé of cryptovolcanic structures, although it consists of an elliptical intrusive complex about 3 miles across and is within the compressional structures of the Ouachita Mountains. See map of Fig. 16.4. The igneous rocks are alkaline, and for the most part belong to the nephelite-syenite group (Landes, 1931). The peripheral intrusives, which are more resistant to erosion than those toward the center of the complex, form a circular ring. Metamorphosed sandstone and shale border the intrusions in places.

Some time after the folding of the Ouachitas, a stock of highly alkalic magma was intruded into the Paleozoic rocks, and then either by differentiation or through separate intrusions several rock types were formed and an unusual suite of minerals was emplaced. Compounds of titanium are especially abundant.

Upheaval Dome

The Upheaval dome is in the flat-lying red-beds of the Colorado Plateau and is sharply conical with a surrounding ring-like syncline. From the axis of the syncline on one side to the axis of the syncline on the other is only 2 miles, and the diameter of the entire affected area is 3 miles. The White Rim sandstone member of the Cutler formation appears as huge, up-ended blocks the size of a house in the highly disturbed central area (McKnight, 1940), and the cliff-making Wingate sandstone rings a spectacular crater about a mile in diameter.

Both aeromagnetic and gravity surveys have been made of the area. The magnetic contours resolve two strong and symmetrical highs, one directly over the Upheaval dome and the other about 7 miles to the southeast. The gravity survey also indicates two structures in about the same places but not so distinctly. Joesting and Plouff (1958) conclude that the broad magnetic and gravity highs each require the rise of a mass of Precambrian crystalline rock about 5 miles in diameter 2000 feet above its normal position. Salt flow emphasized the one dome (Upheaval) but failed to materialize for some unknown reason in the other. Lastly, because the gravity anomalies are not entirely satisfied by the salt plug, a small igneous intrusion into the salt of the dome is postulated. The process took place in several steps from Permian to the Miocene. Refer to "salt anticlines" in Chapter 26 on the Colorado Plateau.

Meteorite Impact Origin

With the space age has come increased interest in terrestrial meteorite impact craters, and Dietz (1960) has called attention to this theory of origin, especially for such structures as Serpent Mound (Fig. 16.3) and the Wells Creek basin (Fig. 16.2). Evidence for the impact theory comes from the presence of shatter cones (small percussion fractures in conical shape) and coesite powder, a high pressure crystalline form of silica, supposedly generated at the time of impact. According to some geologists, the theory is gaining much favor.

17.

MESOZOIC SYSTEMS ALONG THE PACIFIC

WESTERN NEVADA

Central and western Nevada and all California were involved in orogeny during the Mesozoic era, and the index map, Fig. 17.1, shows the chief areas and features with which the following discussion is concerned. The map also extends eastward to central Utah where late Mesozoic disturbances occurred. These will be discussed in Chapter 18.

A trough of geosynclinal proportions centered in western Nevada in Triassic and early Jurassic time. It has already been referred to in connection with the Permian and Mesozoic geanticline in central Nevada. See tectonic maps, Plates 8, 9, and 10. In general, its axis probably lay slightly east of the axis of the Permian trough. In the Hawthorne and Tonapah quadrangles, Nevada, it sank and received a total thickness of sediments of about 30,000 feet (Muller and Ferguson, 1936). The sediments are predominantly marine clastics, cherts, and limestones with a considerable proportion of more or less altered pyroclastic rocks and lavas in the lower and upper parts of the section.

The table, Fig. 17.2, shows the sequence of Mesozoic formation there and elsewhere in western Nevada, California, and southern Oregon. The Lower Triassic Candelaria formation rests with marked erosional unconformity on the thin Permian sandstones and grits and in places on the beveled Ordovician strata. A slight disturbance, therefore, affected the area in late Permian time and probably reflects greater orogeny in the westward-lying orogenic belt. During the deposition of the Candelaria formation, the area of sedimentation as well as the western highland were comparatively quiet, and shales, sandy shales, sandstones, some of tuffaceous aspect, and scattered, thin layers of limestone were deposited. Then marked volcanism and orogeny occurred to the west in middle Triassic time, and over 12,000 feet of strata, chiefly pyroclastics and lavas, accumulated. This group of rocks is known as the Excelsior formation. The lavas range in composition from andesite through quartz latite to rhyolite. Alteration, principally epidotization and chloritization, has affected the formation over wide areas. Volcanic breccias, especially those containing altered andesite fragments, are abundant; and in some sections they exceed the effusive rocks in amount. In the Pilot and Excelsior ranges, a considerable thickness, estimated to exceed 8000 feet, consists of massively bedded chert. Examination under the microscope shows this rock to be an extremely fine-grained water-laid tuff, cemented and largely replaced by cryptocrystalline quartz. Interbedded with the chert are dark tuffaceous slate, a little impure sandstone, and some lava and breccia.

The volcanics were then subjected to erosion for a time but not much disturbed before the thick Upper Triassic sequence accumulated. Dark limestone and dolomite predominate, but siliceous argillite, argillite, calcareous shale, shale, and chert pebble conglomerates are not uncommon. These beds are known as the Luning formation. Above the limestone and dolomite sequence are 420 feet of purple to black shale and dark brown limestone, known as the Gabbs formation. The Gabbs is conforma-

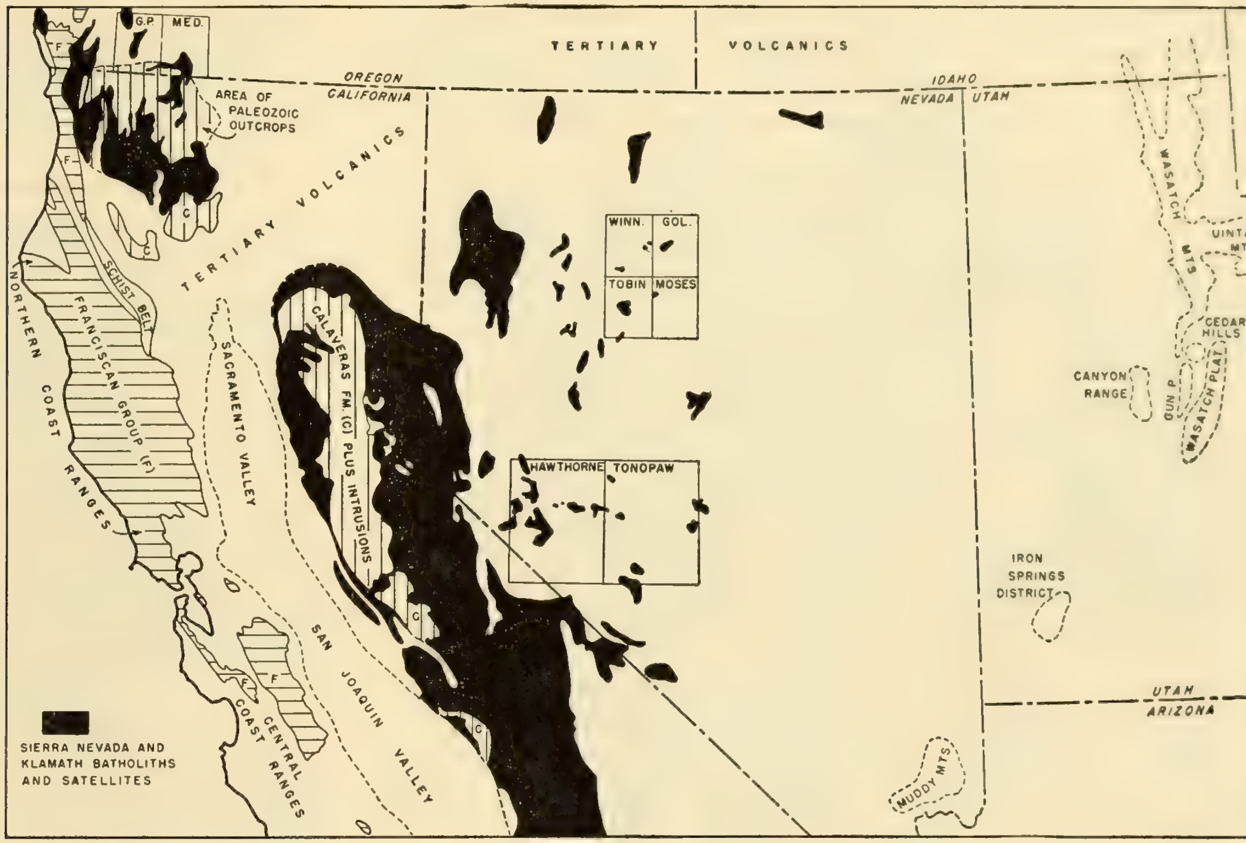


Fig. 17.1. Index map showing significant features and localities of Mesozoic orogeny in the western Cordillera. G.P., Grants Pass quadrangle; Med., Medford quadrangle; Winn., Winnemucca quadrangle; Gol., Golconda quadrangle; Tobin, Mt. Tobin quadrangle; Moses, Mt. Moses quadrangle; Gun. P., Gunnison Plateau.

ble with underlying and overlying formations. Deposition was continuous from Triassic to Jurassic time, while the western orogenic belt remained fairly quiet. Its relief was evidently low, and volcanism is not recorded in the shales, limestones, and sandstones of the Sunrise formation which were deposited in the trough.

At this stage in early Jurassic time, the sediments of the trough were sharply folded (Ferguson and Muller, 1937). The most intense deformation was approximately coextensive with the area of deposition of the Upper Triassic deposits. The orogeny began apparently with the formation of a marginal trough at the border of the geosyncline. In the trough, the Dunlap formation of Early Jurassic age was deposited. It consists dominantly of fanglomerate, conglomerate, and sandstone with an upper volcanic member of andesitic, quartz-latitic, and rhyolitic composition. The fanglomerates and conglomerates were derived chiefly from the limestones of the Luning formation and only locally from the great Excelsior volcanic series. The Dunlap has been observed resting on upturned cherts of the Excelsior formation with an angular discordance of 90 degrees, and also to be truncating folds of the Luning limestones. The Dunlap is characteristically an orogenic deposit, and Ferguson and Muller (1937) recognize a continuation of deformation during its deposition.

These movements were the beginning of thrusting, at least in the area of former deposition. Later compression resulted in thrusting on a large scale, and the earlier structures were greatly complicated. The thrusting postdates the Dunlap Lower Jurassic formation, and precedes the in-

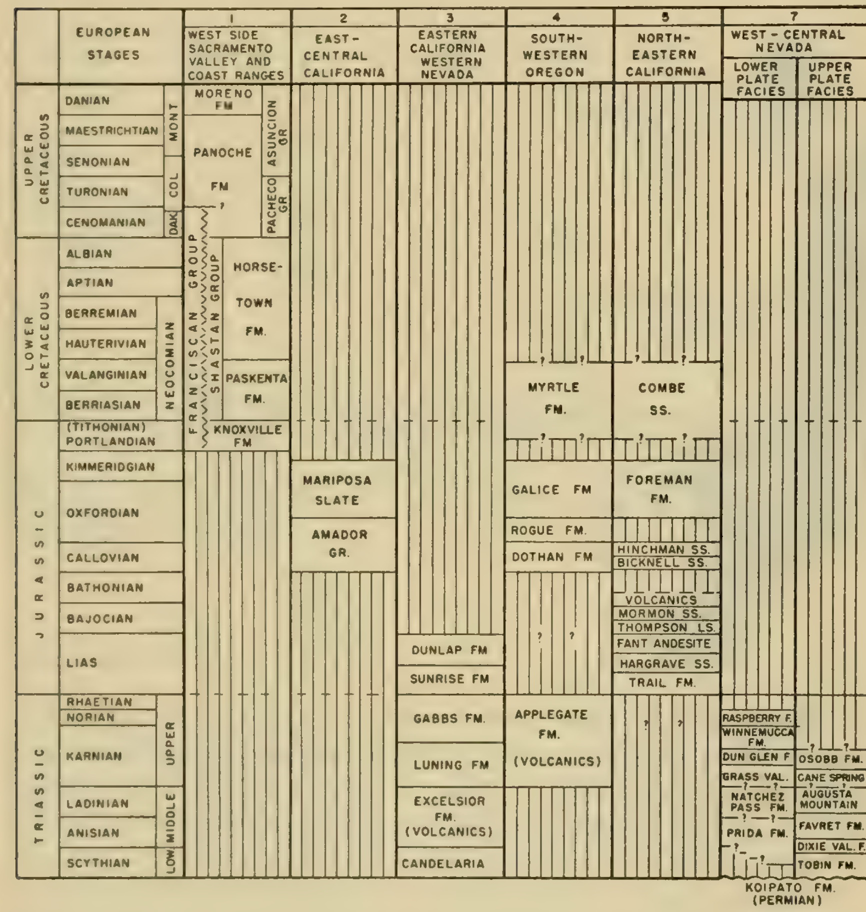


Fig. 17.2. Principal Mesozoic formations of California and western Nevada. West side, Sacramento Valley and Coast Ranges, taken from Irwin (1957) and Briggs (1953). Potassium argon dating of Nevadan orogeny by Evernden *et al.*, 1957. Jurassic correlations from McKee *et al.*, 1956. Triassic of eastern Nevada from Ferguson and Muller (1937), of west-central Nevada (Mount Tobin Quadrangle) by Muller *et al.* (1951), and of southwestern Oregon by Wells (1956).

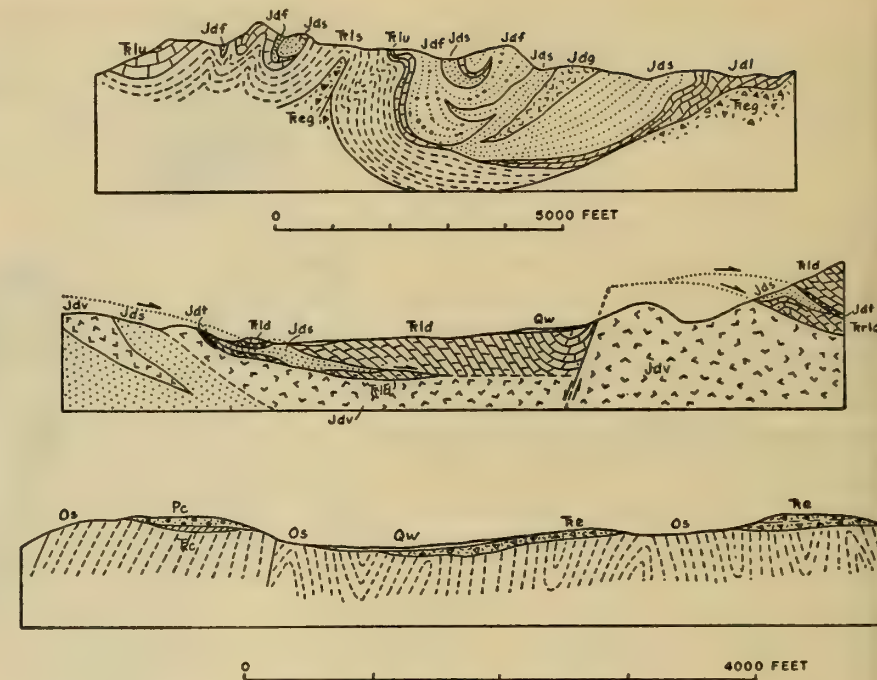


Fig. 17.3. Cross sections in the Hawthorne and Tonopah quadrangles, reproduced from Ferguson and Muller, 1949. Top section, north of Garfield Flat showing relation of Dunlap formation to Luning and Excelsior formations. Middle section, south of Sunrise Flat, Gabbs Valley Range, showing thrusting and later normal faulting. Bottom section, south of Redlich siding, showing relations of Ordovician and Permian and the Excelsior formation. Symbols, top section: Jdf, Dunlap fanglomerate and congl.; Jdg, Dunlap vols.; Jds, Dunlap ss.; Jdl, Dunlap ls.; Ʀlu, Luning upper ls.; Ʀls, Luning slate; Ʀeg, Excelsior vols.; Ʀec, Excelsior chert and tuff. Middle section; Jdv, Dundap vols.; Jds, Dunlap ss.; Jdc, Dunlap congl.; Ʀld, Luning dol.; Jdt, thrust cong. Bottom section: Ʀe, Excelsior vols.; Ʀc, Candelaria formation; Pc, Permian congl.; Os, Ordovician slate and tuff.

trusion of the Sierra Nevada batholith, whose satellites are present in the western part of the sediments of the Triassic and Jurassic trough.

The thrusting in general was easterly along the eastern margin of the trough and southerly along the southern border.

Representative sections from the Hawthorne and Tonopah quadrangles are reproduced in Fig. 17.3, and the evolution of the complex thrust structure in Dunlap and post-Dunlap time is shown in Fig. 17.4.

In the Winnemucca, Golconda, Mt. Tobin, and Mt. Moses quadrangles of west-central Nevada (column 7, Fig. 17.2; area denoted as W–G–T–M on Fig. 17.7) the late Paleozoic Antler orogeny is strikingly displayed, as well as strong orogeny in mid-Permian time. Volcanism in late Permian

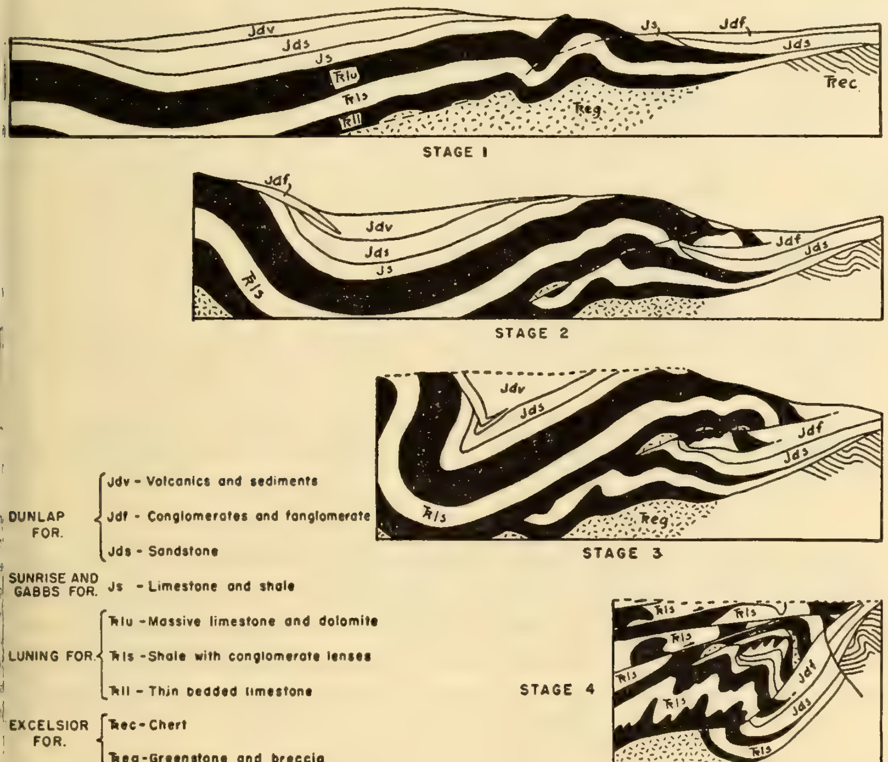


Fig. 17.4. Development of complex structure in the northwestern part of Pilot Mountains, Hawthorne and Tonopah quadrangles, Nev. From Plate 3, Ferguson and Muller, 1949. Stage 1, folding near margin of Luning embayment and deposition of conglomerate and fanglomerate of the Dunlap formation. Stage 2, development of Mac thrust. Deposition of coarse material and folding of Mac thrust. Stage 3, further folding with development of Spearfish thrust. Movement toward the trough was along an erosion surface cut on the upper plate of the Mac thrust. Stage 4, development of five other thrusts and intricate folds. The relative length of the four diagrams indicates the postulated shortening of the stratigraphic section involving the Triassic and Jurassic sediments.

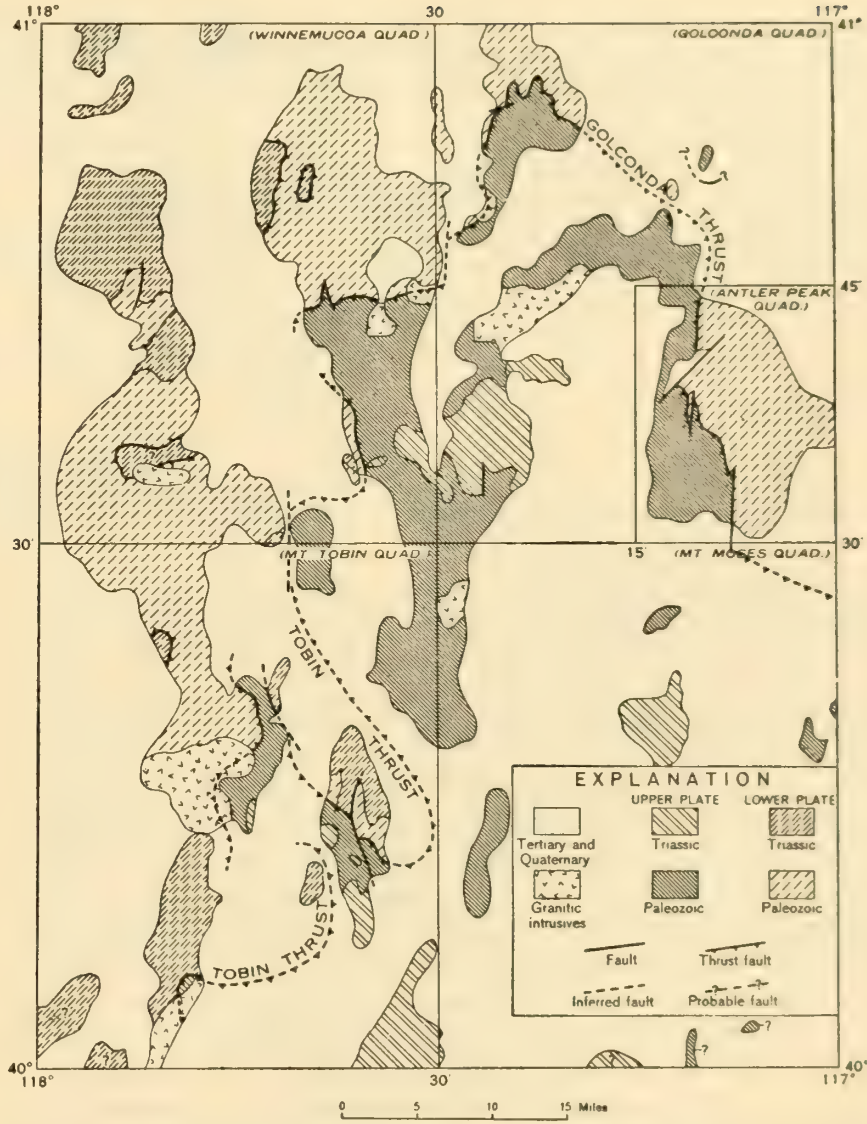


Fig. 17.5. Map showing inferred extent of Tobin and Golconda thrusts. Reproduced from Ferguson et al., 1951.

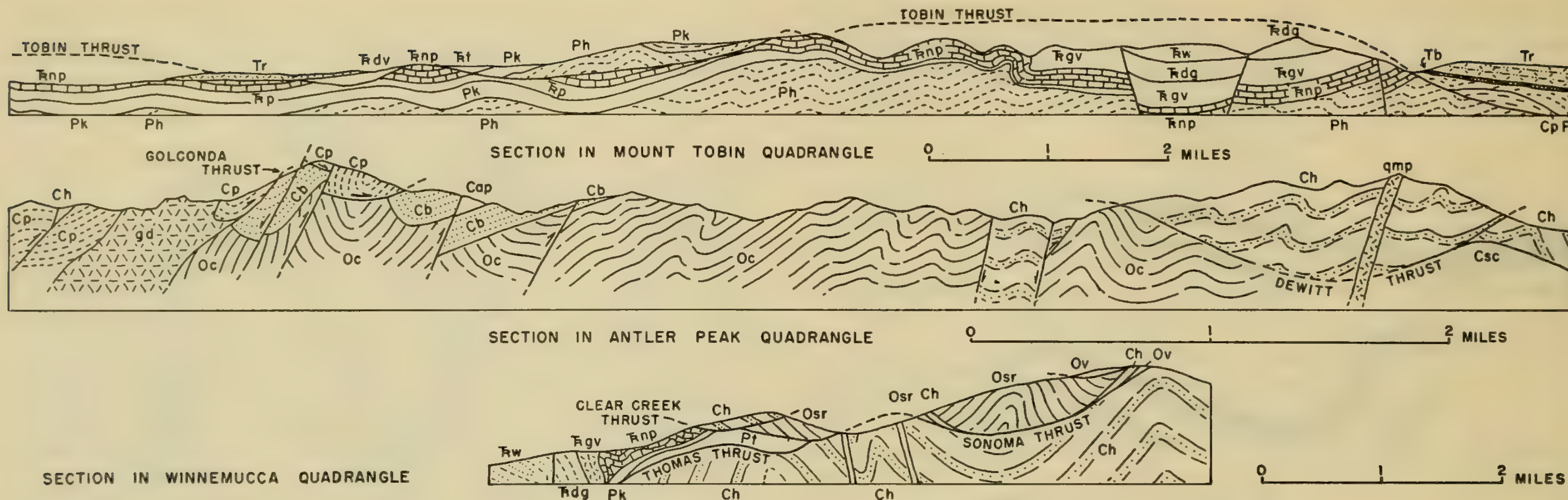


Fig. 17.6. Representative cross sections of northwest central Nevada. Top section shows the Tobin thrust of Late Jurassic age and the angular unconformity between the Permian Koipato and Havallah formations. Middle section shows the Dewitt thrust of late Mississippian or Early Pennsylvanian age and the associated angular unconformity between the Pennsylvanian Battle Mountain formation and the Ordovician Comus formation. Lower section shows the succession of thrusts; first the Thomas, then the Sonoma, and then the Clear Creek, all of Late Jurassic age. The Tobin thrust nearby cuts the Clear Creek thrust.

and, again, mild orogeny at the end of the Permian is noted. See top and middle sections of Fig. 17.6.

Large-scale thrusting occurred in the late Jurassic probably corresponding in time to the major deformation in the Hawthorne and Tonopaw quadrangles. Considering the time of intense deformation of the Mariposa slate in eastern California, to be discussed immediately, the orogeny is thought to have culminated in Kimmeridgian time of the Late Jurassic. The distribution of the major thrusts of this age, the Tobin and Golconda, is shown in Fig. 17.6. The two may actually be one and the same. At least, the horizontal translation has been so great that two suites of formations of different facies probably deposited an appreciable distance apart, have been brought into juxtaposition. In the four quadrangles the upper thrust plate covers an area extending 50 miles from north to south and 40 miles from east to west. The Permian formations are common to both plates. The direction of relative movement of the upper plate is uncertain. In the Sonoma Range a succession of four thrusts, all occurring in the Late (?) Jurassic orogeny, is recognized, and the three lower ones moved from east to west. It seems possible that the Tobin thrust plate could have moved toward the north (Ferguson *et al.*, 1951). See lower section of Fig. 17.6.

NORTHWESTERN NEVADA

Lower and probably Upper Cretaceous rocks have been found in northwestern Nevada, and these record a continuation of deformational phases beyond the Late Jurassic Tobin and Golconda thrusting. According to Willden (1958):

TERTIARY		Folding and faulting in northwestern Nevada.	Laramide orogeny?
UPPER CRETACEOUS	Danian	Thrusting of Permian volcanics over King Lear and Pansy Lee clastics.	Late Nevadan orogeny
	Maestrichtian	Deposition of Pansy Lee conglomerate in northwestern Nevada.	
	Senonian	Santa Lucian phase in Central Coast Ranges.	
	Turonian	Intrusion of great batholiths of Sierra Nevada and Coast Ranges.	
	Cenomanian		
	95–101		
LOWER CRETACEOUS	Albian	Folding and erosion of King Lear fm.	Mid-Nevadan orogeny
	Aptian		
	Barremian	Uplift and deposition of King Lear fm. in northwest Nevada.	
	Hauterivian		
	Valanginian		
	Berriasian		
	133		
JURASSIC	Portlandian (Tithonian)	Intrusion of batholiths in southern Klamath Mts. and northwestern foothills of Sierra Nevada.	Early Nevadan orogeny
	Kimmeridgian	Strong orogeny; Tobin and related thrusts of W-G-T-M Quadrangles. Mariposa slate of eastern California isoclinally folded with resulting low-grade metamorphism.	
	Oxfordian		
	Callovian		
	Bathonian	Volcanism and local folding and thrust-faulting during deposition of Dunlap fm. in Hawthorne and Tonopah Quadrangles.	Unnamed phases of orogeny
	Bajocian		
	Lias		
	175		
TRIASSIC	Rhaetian		
	Norian	Subsidence of Luning Embayment.	
	Karnian	Strong, local orogeny and volcanism; folding and thrusting in Hawthorne and Tonopaw Quadrangles.	
	Ladinian		
	Anisian	Mild, local disturbance resulting in angular unconformity.	
	Scythian		
	185–200	Mild orogeny in central and western Nevada resulting in unconformity.	
PERMIAN	Ochoa	Volcanism, extensive. Folding in central Oregon.	
	Guadalupe		Sonoma orogeny
	Leonard	Orogeny in Western Nevada: Golconda thrust	
	Wolfcamp		
	210		
PENNSYLVANIAN	Virgil		Antler orogeny
	Missouri		
	Des Moines	Strong orogeny, folding and thrusting in central and western Nevada. Sharp folding and low-grade metamorphism of Calaveras fm. in eastern California possibly at this time.	
	Lampassas		
	Morrow		
MISSISSIPPIAN		Continued orogeny probably in several phases.	
DEVONIAN		Beginning of geanticlinal uplift in central Nevada, and compressional orogeny in part.	

Fig. 17.7. Sequence of disturbances in central and western Nevada and California from the central Coast Ranges to the Sierra Nevada. Numbers are absolute ages in terms of millions of years and in part are modifications of the Holmes time scale as proposed by Curtis *et al.*, 1958.

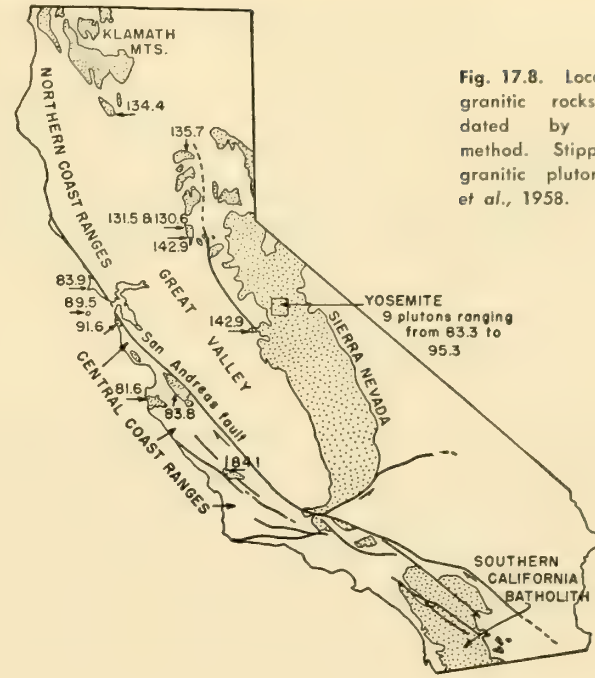


Fig. 17.8. Location and age of granitic rocks in California dated by potassium-argon method. Stippled areas are granitic plutons. After Curtis *et al.*, 1958.

A formation of Early Cretaceous age composed of locally derived clastic rocks, including pebble to boulder conglomerate, siltstone, coarse graywacke, and finely crystalline limestone is exposed at several places in the central and northern part of the Jackson Mountains, Humboldt County, Nevada. This formation (King Lear) was folded and at two places probably completely eroded before deposition of the next younger unit—a pebble conglomerate composed of exotic pebbles of chert and quartzite derived from rocks of early Paleozoic age. This younger pebble conglomerate (Pansy Lee) may be of Late Cretaceous or early Tertiary age and may be equivalent to rocks of similar stratigraphic position and lithologic character exposed over a considerable area of eastern Nevada and western Utah. Both of these coarse clastic formations have been overridden by a thrust sheet of Permian or older volcanic rocks. The dimensions of the thrust sheet are not known exactly but remnants are exposed over a 25-mile-long segment of the range. Upper Tertiary volcanic rocks exposed in the range are not involved in the thrusting.

The Cretaceous and younger rocks of the Jackson Mountains record a long period of orogenic unrest that included: (1) uplift of the source area of and deposition of the Lower Cretaceous rocks; (2) folding and beveling by erosion; (3) deposition of the exotic-pebble conglomerate; (4) thrusting of the Per-

mian or older volcanic rocks over the two coarse clastic formations; and (5) later folding, faulting, and erosion providing the present outline of the range.

These relations record orogeny of Laramide age and undoubtedly mean that the Laramide belt of the Central Rockies (Chapter 22) spread westward over most of Nevada. The folding of the King Lear clastics before the deposition of the Pansy Lee conglomerate records deformation probably in Early Cretaceous time, and this has been labeled the mid-Nevadan orogeny on Fig. 17.7.

CENTRAL AND NORTHERN CALIFORNIA

In the California Sierra Nevada region, Taliaferro (1942) summarizes an eastern belt of Triassic and Jurassic rocks and a western belt of Jurassic rocks. The two were probably continuous, but due to the Nevadan orogeny (to be described immediately) a dividing mass 25 to 50 miles wide of Calaveras rocks and granite of the Sierra Nevada batholith exists. The eastern belt consists of discontinuous areas of Upper Triassic and Jurassic sediments and volcanics. Doubtless these formed a continuous belt at one time, but as they lie in the region of maximum plutonic invasion and maximum Tertiary uplift, they have been obliterated or removed by erosion in many places. Near the northern end of the Sierra Nevada, the Milton formation represents the Triassic and Jurassic rocks, and where not engulfed by the plutons or removed by erosion it lies in a broad, steep-limbed syncline, practically free from minor crumbling, thrusting, and overturning (Taliaferro, 1942). The conglomerates contain abundant debris of the Paleozoic rocks (Calaveras) and thicken and coarsen westward. It seems clear, therefore, that the Milton was derived from the west.

The best-exposed and most complete section of the east belt is on the north fork of the American River in Placer County. On the west limb of the syncline, basic and intermediate volcanics and radiolarian cherts, 200 feet thick, are followed by 12,800 feet of conglomerates, sandstones, hard slaty shales, and fine-grained andesitic tuffs. The center of the syncline is occupied by 9500 feet of intermediate and basic flows, agglomerates, and tuffs. Only about 900 feet of sediments and tuffs lie below the volcanics on the east limb of the syncline, the lower part having been obliterated by batholithic intrusions. Well-preserved Upper Triassic fossils are found at and near the base of the sediments on the west limb of the syncline, Lower Jurassic fossils 2500 feet above the base, and Middle Jurassic fossils 9500 feet above the base; no fossils have been found in the upper 13,000 feet of the sediments and volcanics. The section is well exposed and no unconformities or disconformities have been observed. Possibly part of the upper 13,000 feet is equivalent to the Mariposa slate of the western belt. The upper volcanics are possibly equivalent to the extensive Logtown Ridge volcanics lying between Amador and Mariposa west of the Mother Lode. No evidence supports the idea that the Milton of the eastern belt was separated from the Mariposa and Logtown Ridge of the western belt either by deposition in separate basins or by a period of batholithic intrusion and orogeny (Taliaferro, 1942). See column 2, Fig. 17.2.

In comparing the sediments of the eastern belt of the Sierras with those of the trough of western Nevada, it appears that Lower and Middle Triassic sediments were deposited in the central part of the trough which lay in western Nevada, and then Upper Triassic sediments overlapped on highlands both westward and eastward. See Fig. 17.8. Great subsidence occurred in early Middle, and early Late Jurassic time; the center of the Jurassic trough migrated west of that of the Triassic trough; and overlap on the western volcanic orogenic belt was extensive.

The western belt is made up of the Amador group and the Mariposa slates in the Sierra Nevada and northwestward in Oregon, of the Dothan and Galice. The Amador and Dothan are probably Middle Jurassic in age, with their upper beds containing Late Jurassic fossils. The Mariposa and Galice are early Late Jurassic. The great bulk of the Amador consists of volcanics and clastics, but red and green radiolarian cherts and dense, unfossiliferous limestones are common. On the Cosumnes River, 1200 feet of conglomerates and sandstones are at the base of the Amador. On the Merced River, radiolarian cherts, tuffs, and shales are over 1500 feet thick, and these overlie about 1400 feet of pillow basalts. The entire Amador group ranges in thickness from 5000 to 15,000 feet, and usually

grades upward into the Mariposa (Taliaferro, 1942), but between the Merced and Mariposa rivers, conglomerates are at the base of the Mariposa. The pebbles are presumably from the underlying Amador. See column 4, Fig. 17.2.

The Mariposa formation consists of black slate and graywacke, with which greenstone is closely associated (Knopf, 1929). Conglomerate occurs locally, and sericite schist and limestone in a very few places. The greenstone, because of its intimate interbedding with the normal sedimentary rocks, is in many places an inseparable part of the formation, and locally predominates in volume. The conglomerate contains a variety of rocks, namely: quartz keratophyre (submarine lava flow origin), quartzite, chert, quartz, aplite, and biotite granophyre. The last two point to plutonic intrusions older than those of the Sierra Nevada (Knopf, 1929). The graywacke contains grains of quartz, plagioclase, slate, quartzite, and keratophyre (?). On the one hand they grade into slate and graywacke slate, and on the other, by the presence of augite, into augite tuff. The greenstones were principally augite basalt breccias, tuffs, and lavas, now somewhat metamorphosed (Knopf, 1929). It appears that some of the volcanics included by Knopf in the Mariposa are what Taliaferro places in the Amador.

The great thickness of volcanics is a striking feature of practically all Jurassic units in California and southwestern Oregon. The volcanic rocks range from rhyolite to basalt, but augite andesites predominate. Practically all, if not all, are submarine, as they are interbedded with marine sediments (Taliaferro, 1942). Pyroclastics predominate over flows. Centers of volcanism have been recognized in the form of necks, both breccia and solid, and great accumulations of flows, tuffs, and very coarse breccias.

Intrusions of peridotite and dunite, now largely serpentinized, together with their closely associated gabbroic and diabasic differentiates, are common in the Jurassic of California and southwestern Oregon. They occur as sills, dikes, plugs, and large masses of undetermined form. The great majority were intruded prior to folding of the Jurassic sediments and before the Sierra Nevadan batholith was emplaced. The basic intrusions of the Mother Lode were serpentinized immediately after their emplacement (Knopf, 1929). They were slightly metamorphosed by the folding, and greatly altered at the contacts of the granodiorite plutons.

OREGON

In central Oregon, a fairly complete Jurassic section has been described by Lupher (1941). He sets apart ten formations which range in age from Early to Late Jurassic, perhaps to Early Cretaceous, and altogether are over 11,000 feet thick. These beds show only a succession of gentle emergent and submergent movements. The lithology is in conspicuous contrast to that of the Jurassic of the Sierra Nevada in lacking volcanics and having only minor amounts of coarse clastics. It is nearly all sandstone and shale, and in part it is very fossiliferous.

The Oregon Jurassic rests with marked angular discordance on a basement of highly folded Upper Triassic and Mississippian rocks. Some of the beds called Upper Triassic may be Lower Jurassic, because a sequence of shales, sandstones, and conglomerates, many thousands of feet thick, overlies the fossiliferous Upper Triassic but underlies the great unconformity. The folds in the Jurassic beds trend at divergent angles from those of the Upper Triassic, and basic plutons now largely altered to serpentine invade the Upper Triassic but not the Jurassic. It is, therefore, apparent that an orogeny of considerable proportions is indicated. It will be recalled that a similar unconformity separates two formations of Early Jurassic age in western Nevada, and it is evident, therefore, that the two may be the same, perhaps with slightly different ages. It seems necessary, in order to account for the different lithologies of the Jurassic beds of central Oregon and those of the Sierra Nevada, to separate the central Oregon beds from the volcanic belt by a nonvolcanic highland or Piedmont. See the tectonic maps, Plates 11 and 12. The pebbles are cherts and limestones, evidently from Paleozoic formations (Lupher, 1941).

SOUTHERN CALIFORNIA

Larsen (1948) has summarized the geology of the region southeast of Los Angeles in southern California, especially in relation to the great

Nevadan intrusions there; and he also reviews the southern continuation of the batholithic province into Baja California. A group of slates and argillites, with some quartzites, lie west of the main batholith and form most of the Santa Ana Mountains. Triassic fossils have been collected there in several places. Somewhat more metamorphosed remnants of these rocks occur within the batholith. The group is known as the Bedford Canyon formation, and about 20,000 feet of beds are exposed in the Santa Ana Mountains. Parts of the formation may be older than Triassic and parts may be Jurassic. The uniform argillaceous lithology is a dominant character.

A group of volcanic beds, mostly mildly metamorphosed, andesitic agglomerates, overlies the Bedford Canyon formation unconformably. The extrusives have been called the Santiago Peak volcanics (Larsen, 1948), and they are probably many thousands of feet thick. They are older than the batholithic intrusions and, therefore, are probably Jurassic in age.

Along the east side of the main batholithic region are coarsely crystalline schists, all of which contain much quartz. Interbeds of limestone have yielded Mississippian fossils (Larsen, 1948). A quartzite sequence with interbedded, coarse, mica schist is also thought to be Carboniferous. It is some 12,000 feet thick. Larsen believes that the Paleozoic sediments were metamorphosed and intruded by granite rocks before the deposition of the Triassic rocks, and that this older metamorphism was more intense than the later metamorphism of the Triassic rocks.

NEVADAN OROGENY

History of Concept

The literature, up to the last few years, suggests that the Late Jurassic folding and thrusting was followed very shortly by the great batholithic intrusions, and that the two events occurred between the Kimmeridgian and Portlandian. See Figs. 17.2 and 17.7. Recent isotope age determinations have demonstrated fairly conclusively, however, that the intrusions are mid- or early Late Cretaceous in age. Also new fossil finds have resulted in a revision of concepts of the Upper Jurassic and Lower Cretaceous stratigraphy which is not incompatible with a Mid-Cretaceous age of the batholiths.

Additional sampling and potassium-argon age determinations by Curtis *et al.* (1958) indicate that granitic rocks along the northwest foothills of the Sierra Nevada and in the southern Klamath Mountains are Tithonian (Portlandian) in age, as the early geologists had concluded. Furthermore, they found that several plutons in the Central Coast Ranges are early Late Cretaceous (about Cenomanian to Senonian), the same age as the plutons of Yosemite National Park. The various potassium-argon ages to date in California are shown in Fig. 17.8. Curtis *et al.* conclude that the bulk of the great batholiths of California are of the later date, but that some are late Jurassic, and, as the earlier writers concluded, are closely associated with the post-Kimmeridgian folding and thrusting.

The term Nevadan orogeny has been used to denote those tectonic events that occurred in the general region of the Sierra Nevada in a rather limited interval of time between the Kimmeridgian and Portlandian. The great batholiths are indelibly impressed in the literature as an outstanding characteristic of the orogeny, so now with the recognition that the main batholiths are much younger we are faced with a redefinition of the term, Nevadan orogeny. It is here proposed to call those disturbances and intrusions in Late Jurassic time (post-Bathonian) the early Nevadan orogeny, those of Early Cretaceous time the mid-Nevadan orogeny, and those of Mid- and early Late Cretaceous time the late Nevadan orogeny (see Fig. 17.7).

General Characteristics

The Jurassic and pre-Jurassic rocks thus far described were severely folded and thrust-faulted in the Sierra Nevada, and then invaded by granitic magma. The maximum deformation seems to have been concentrated along what is now the western slopes of the Sierra Nevada in the zone of the western belt of Jurassic deposits. Overturned folds, some of great amplitude, great thrusts, such as the Mother Lode zone, and mild dynamic metamorphism were widespread. The eastern belt of Triassic and Jurassic rocks, near the present crest of the Sierra Nevada, is strongly folded, but less dynamically metamorphosed. The eastern belt of Triassic

and Jurassic rocks continued southward into southern California, but is much obscured there by Tertiary lavas and late Cenozoic faulting. See Fig. 17.9.

Central and Northern California

At the north end, in the Taylorsville region, the Paleozoic rocks are thrust eastward over the Jurassic, overturning them toward the east, just the opposite of the thrusting along the Mother Lode on the west flank of the Sierra Nevada. It will be recalled that the Late (?) Jurassic thrusting in western Nevada was both toward the east and west and locally probably southward and northward. In the Grass Valley area of the northern Sierra Nevada, Johnston (1940) finds the rocks to have been compressed into northwest-trending isoclinal folds. The metamorphism was of the feeblest kind. The Mariposa was compacted and cemented, and some of the andesitic rocks acquired schistocity; but the chemical and physical changes were much less severe than those imposed upon the Calaveras formation in late Paleozoic orogeny.

Regarding the post-Mariposa plutons, Knopf (1929) says that in the Mother Lode belt the oldest of these rocks there are peridotites which, soon after intrusion, were transformed into serpentines. Smaller masses of gabbro and hornblendite were then intruded into the peridotite, to which they seem to have a predilection. The peridotite, gabbro, and hornblende appear to represent the "basic prelude" to tremendous intrusions of granodiorite that form the bulk of the present Sierra Nevada.

The granodiorite is uniform in texture and composition, and contains basic clots which are very common in the high Sierra. Quartz diorite porphyry is intrusive into the Mother Lode belt south of Placerville. It grades into the granodiorite and has exerted no perceptible contact metamorphism. Knopf believes that the granodiorite ascended to a high level in the earth's crust in the gold belt area. Dikes and small intrusive masses of a white rock composed almost entirely of albite complete the intrusive cycle. Allied varieties of the granodiorite are quartz-monzonite, granite, and alaskite. The Mariposa is affected by contact metamorphism as much as a mile away from the granodiorite contact.

In the northern Sierra Nevada, Johnston (1940) finds essentially the same batholitic cycle as Knopf does to the south in the Mother Lode, namely, an intrusive succession of ultrabasic rocks, gabbro, diabase, granodiorite, granite, and aplite. Granodiorite was intruded in tremendous batholithic masses that now form the backbone of the high Sierra. On the western slope, smaller masses of granodiorite are satellitic to the main mass. The earlier formations were shoved aside and possibly in part assimilated, and contact metamorphic zones were developed in the sedimentary rocks. From the last emanations of the granitic intrusions were formed the gold quartz veins of the Sierra Nevada.

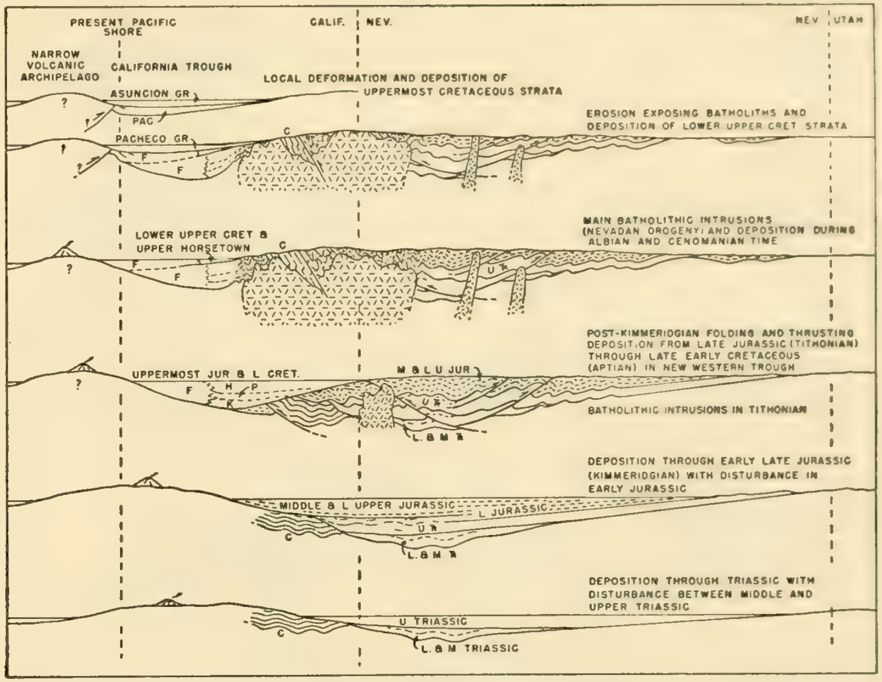


Fig. 17.9. Evolution of the Sierra Nevada through Mesozoic time. C is Calaveras formation of late Mississippian (?) age; F is Franciscan group; K, Knoxville formation; P, Pashenta formation; and H, Horsetown formation.

In the southern part of the Sierra Nevada, Mayo (1941) in reviewing the work of others and himself, finds that hornblende gabbro and hornblende diorite were forerunners to the main granitic intrusions. These

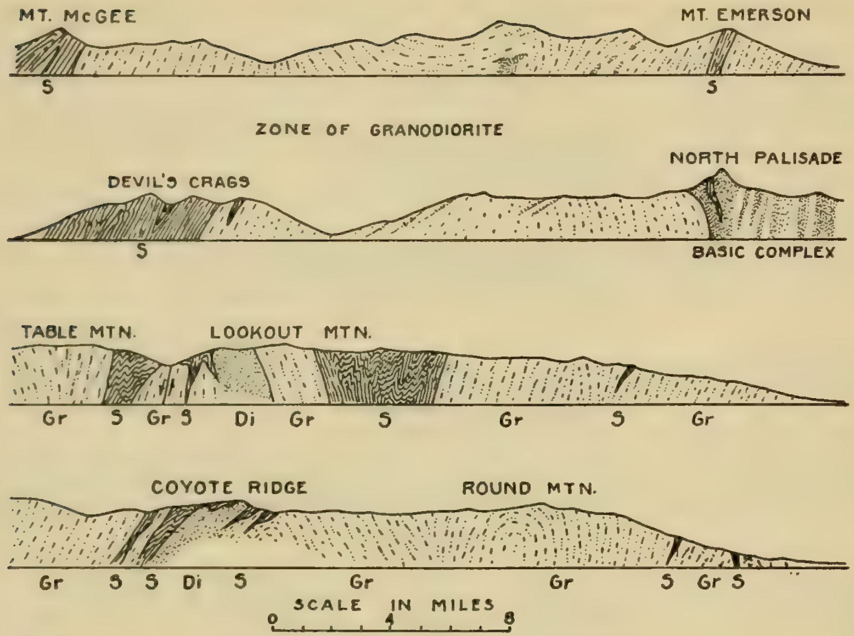


Fig. 17.10. Structure sections across southern Sierra Nevada Mountains. Upper section is north of Connecting Link; second section is south of it. Lower two sections are across the northern part of Coyote Salient. s, septum; Gr, granitic rocks; Di, Diorite and gabbro. After Mayo, 1941.

basic intrusions now appear widely distributed as dark zones, strips, and masses of various shapes and simulate the remnants of the metamorphic rocks. According to Mayo, the bulk of the Sierra Nevada core ranges in composition from granodiorite to granite, with quartz monzonite predominating. All members of the intrusive sequence are penetrated by dikes of aplite and pegmatite. Some basic dikes were late comers also.

The groups of intrusions are separated at many places by long, narrow strips and by local broad areas of metamorphic rocks. The metamorphic rocks are divisible into two groups: an older series of metasediments of probable Paleozoic age, and a series of metavolcanics, part of which Knopf has assigned to the Triassic.

The metamorphic rocks are remnants of septa (Fig. 17.10) that divided the intrusions to unknown depths. During the earliest recorded deformation, the original bedding and other layered structures were thrown into a series of closely appressed, nearly vertical-sided, isoclinal folds. Cleavage developed approximately parallel to the axial planes of the folds, and was followed by many small shears and a few upthrusts. Linear structures that vary greatly in pitch were formed in the planes of cleavage, bedding, shears, and upthrusts. These metamorphic rock structures are separated from the intrusions by contacts that are usually very steep and sharp. Gradational contacts are suggested in a few places.

Within the granitic rocks, a parallel arrangement of inclusions, minerals, and schlieren reveals layered and linear traces of flow that are assigned to the plastic stage of intrusion. These structures of the plastic stage, by grading into fractures, locally record the stage of transition. The stage of transition was followed by the solid stage, when adjustments resulted in fracturing.

In the Huntington Lake area of the western slope of the central Sierra Nevada, Hamilton (1956a) has concluded that the crystalline rocks there consist of ten separate, sharply bounded, plutons which range in size from one square mile, approximately, to several hundred square miles. Only small parts of this area consist of metamorphic rocks. See Fig. 17.11.

The granite rocks range from alaskite to quartz diorite, but it is important to note that a rock type does not constitute a separate intrusion, but rather, each intrusion may be made up of two or more rock types. Two of the plutons range from quartz diorite through granodiorite to quartz monzonite. In another, the content of ferromagnesian minerals varies from 2 to 19 percent. The abundance of ferromagnesian minerals and of the dark inclusions are closely parallel. The inclusions are xenolithic, and some and possibly most of the mafic minerals are products of assimilation of metamorphic rocks. Most of the granite rocks are believed to have formed from the upward intrusion of mobile materials.

The western group, consisting of the Tamarack Creek, Huntington Lake, Sheepthief Creek, and Kaiser Peak plutons, is considered the older, and eastern group, consisting of the Mt. Givens, Red Lake, Rodeo Meadow, Dinkey Lake, Coyote Creek, and Helms Creek plutons, the younger.

The relative aerial abundances of the rock types are as follows:

alaskite	5	percent
granite	4	percent
quartz monzonite	47	percent
granodiorite	33	percent
quartz diorite	11	percent

This confirms Mayo's observation that quartz monzonite is the most voluminous rock type in the Sierra Nevada where studied petrographically.

Age of the Batholiths

The first determination of the age of the Sierra Nevada batholith by isotope methods was made by Larson *et al.* in 1954. Lead-alpha activity ratios were determined on the accessory minerals zircon, monozite, and xenotime. Seven samples yielded an average age of 100 m.y. Twenty-five samples were run from the batholith of southern California, and these gave an average age of 105 m.y. (Larson *et al.*, 1954).

A few years later samples were taken by Evernden *et al.* (1957) from eight individual intrusions in the Yosemite Canyon area of the Sierra Nevada batholithic complex, plus a pegmatite of one of the plutons and their ages determined by the potassium-argon method. The major plutonic bodies had been mapped by Calkins (1930) and Rose (1957) who had established for the most part the relative ages of the intrusions on convincing field evidence. From youngest to oldest the seven plutons are named as follows: Johnson granite porphyry, Cathedral Peak granite, Half Done quartz monzonite, Sentinel granodiorite, El Capitan granite, Gateway granodiorite, and Arch Rock granite. The Hoffman quartz monzonite, which is noted to have intrusive relations to the Cathedral Peak granite, was also sampled. The experimental age determinations agreed perfectly with the relative ages determined by geological field relations. The youngest, the Johnson granite porphyry, yielded a date of 82.4 (±1–2%) m.y., and the oldest, the Arch Rock granite, 95.3 (±1–2%). The authors from theoretical considerations regard these ages as slightly younger than the true absolute ages of the plutons, but believe any change made ultimately will be in the order of a few percent at most.

A pegmatite in the Hoffman pluton (83.3±1–2% m.y.) yielded an age of 76.9 m.y., and the range from this youngest rock to the oldest is therefore approximately 18 m.y. This intrusive activity would have occurred according to Curtis *et al.* (1958) during the Cenomanian, Turonian, and Senonian (see Fig. 17.7) epochs.

If the series of nine plutons, including a late pegmatite, were intruded during an interval of 18 m.y., a separate intrusion approximately each 2 m.y. would have been emplaced. Evernden *et al.* (1957) review the field evidence to the effect that most of these intrusions were almost completely crystallized at the time the succeeding pluton was emplaced, and thus conclude that crystallization of each would require somewhat less than 2 m.y.

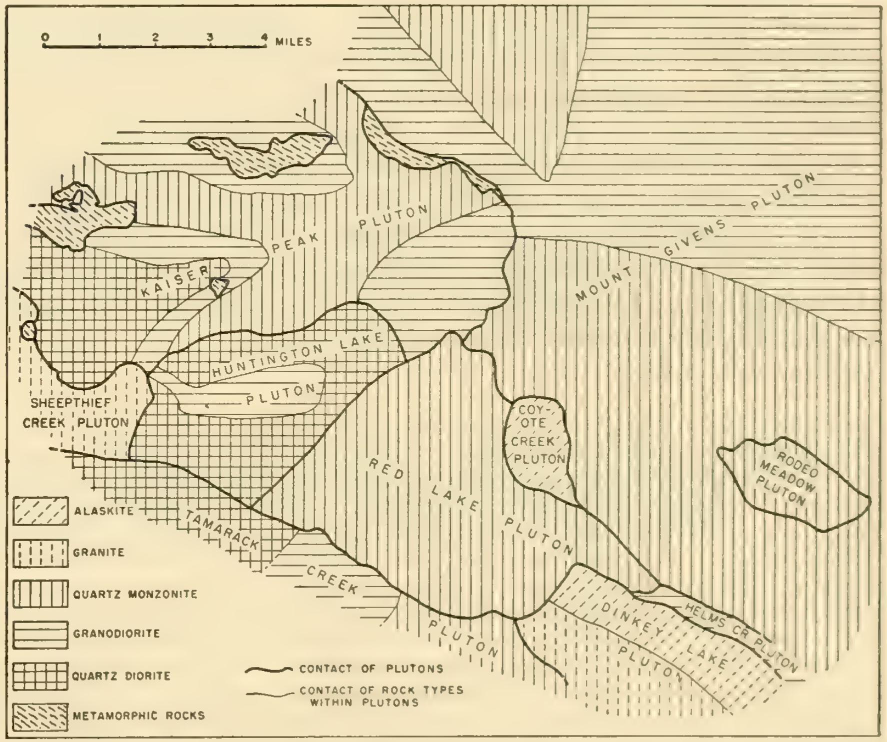


Fig. 17.11. Plutons and rock types of the Huntington Lake area; Sierra Nevada batholithic complex. Direction of pattern lines has no significance.

The granitic rocks were exposed by erosion at the time the Turonian sediments were deposited (see Figs. 17.7 and 17.8) and hence, only a short time separated the last intrusion from its exposing. It may thus be assumed that uplift and erosion kept close pace with granitic emplacement. From this Evernden *et al.* deduce that the space for the batholiths was produced by the elevation of the roof slowly and by small increments, and that the overlying sedimentary rocks were stripped by erosion as rapidly as they rose.

ANCESTRAL COAST RANGE SYSTEM

Franciscan Basin

Following the post-Kimmeridgian folding and thrusting (Fig. 17.9) a trough or basin sank on the west in California, and in it an exceedingly thick sequence of sediments accumulated. These are those of the Franciscan group and equivalents (Fig. 17.2). West of this trough lay a sourceland of sediments, viewed as a narrow volcanic archipelago by Taliaferro (1942). The strata known as Franciscan crop out in the Coast Ranges and the Shasta and Upper Cretaceous strata occur in the Sacramento Valley. The thickness of the Franciscan is about 35,000 feet. That of the Shasta series is about 10,000 feet and of the Upper Cretaceous on the west side of Sacramento Valley is 15,000 feet.

According to Irwin (1957):

The Franciscan group consists dominantly of detrital sedimentary rocks with interbedded chemical sedimentary and volcanic rocks. The detrital rocks are chiefly sandstones of the graywacke type, with interbedded shale and conglomerate. Reliable criteria have not yet been described for distinguishing, either in hand specimen or under the microscope, between detrital rocks of the Franciscan group and those of the Sacramento Valley sequence. The most obvious and significant difference between the lithologic character of the Franciscan group and that of the Sacramento Valley sequence is the presence and local abundance of interbedded volcanic rocks and associated chemical sedimentary rocks in the Franciscan. The chemical sedimentary rocks include rhythmically thin-bedded chert, and, much less abundantly, a distinctive foraminiferal limestone. In addition, the Franciscan group includes small areas of glaucophane schists. In some areas, strata of the Franciscan group have been metamorphosed to slates and phyllites.

The Franciscan group has been intruded by mafic and serpentinized ultramafic rocks, and has been highly faulted and pervasively sheared. The general appearance of the Franciscan terrane, because of the net effect of the lithologic heterogeneity and complex structural deformity, is in striking contrast to areas underlain by strata of the Sacramento Valley sequence.

The Knoxville formation as exposed along the west side of Sacramento Valley between Wilbur Springs and Paskenta is perhaps 10,000 feet in average thickness. The base is unknown, as along most of the valley the lowest exposed beds are in fault contact with the belt of ultramafic rock. The Knoxville formation is generally considered to consist typically of a thick section of thin-bedded shales with small lenses of limestone, but interbedded sandstones and conglomerates are locally abundant. Fossils indicate that it is Late Jurassic (Tithonian) in age. One of its most characteristic and abundant fossils is Aucella piochii Gabb.

The contact between the Knoxville formation and overlying Shasta series is marked by a fairly abrupt and complete change in fauna, and at many places by beds of conglomerate. Here, as well as at other places, the concept of a "basal conglomerate" has much influenced the subdivision of the Sacramento Valley sequence. Along much of Sacramento Valley the transition from one unit to the other is one of nearly continuous deposition and, judged from broad structural conformity, was accomplished with little disturbance.

The strata referred to the Shasta series have a higher ratio of sandstone to shale than has the Knoxville formation.

Upper Cretaceous strata along the west side of Sacramento Valley consist of sandstones and shales and are about 15,000 feet in average thickness. They represent only the lower part of the Upper Cretaceous section of northern California.

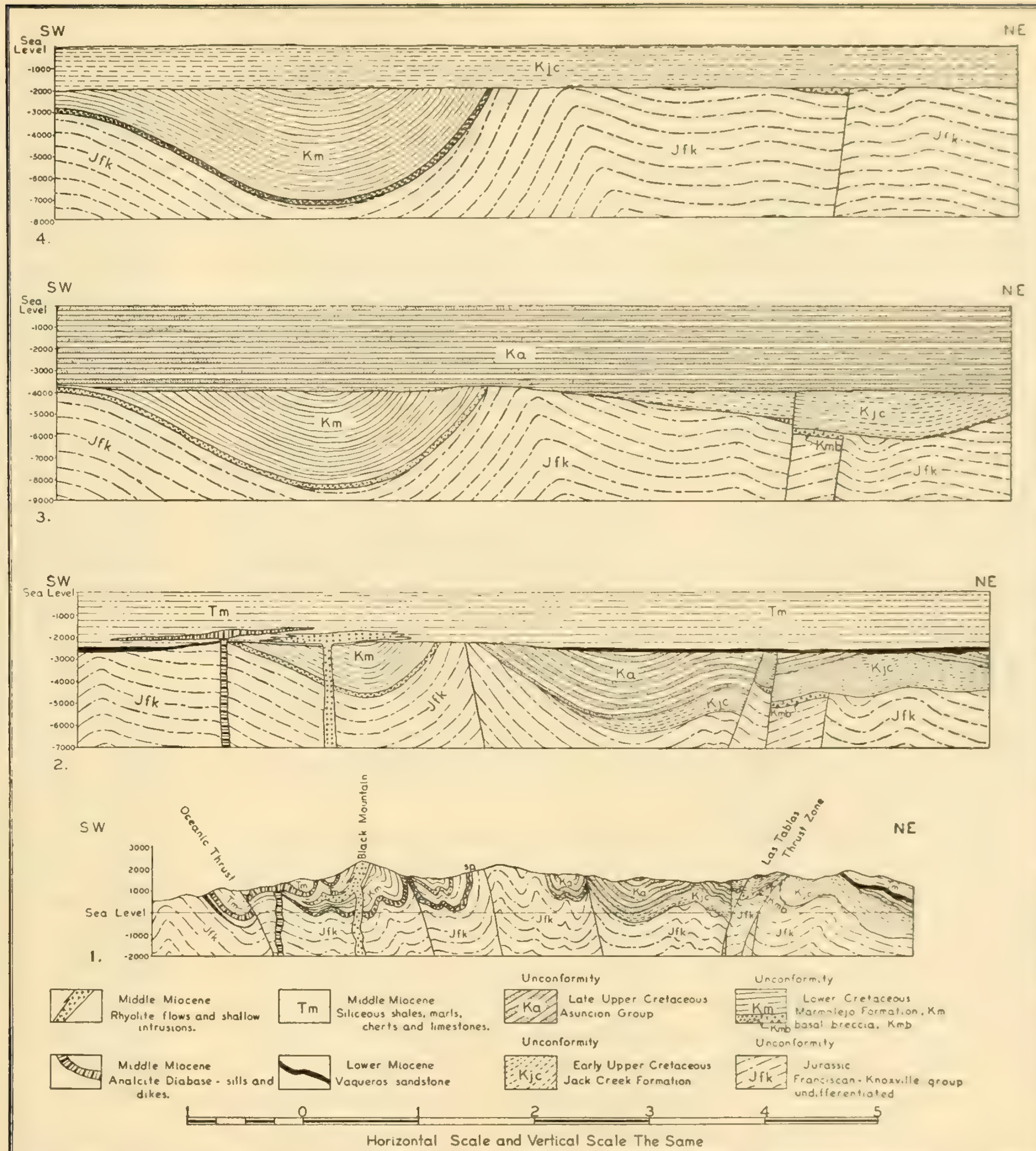
Mid-Cretaceous Phase (Mid-Cretaceous Orogeny)

In many places in the Coast Ranges there is either a definite disconformity or a strong unconformity or overlap between the Shasta series and the Upper Cretaceous strata. Especially in the Santa Lucian Range, there is evidence of deep erosion and overlap. Along the crests of some of the folds produced during this disturbance, the Lower Cretaceous and Upper Jurassic beds were removed, so that the Shasta trough was lifted in subparallel fragments. Other parts of the Shasta beds were little affected. The orogeny represented by the unconformity has been called the Mid-Cretaceous by Taliaferro (1943b).

Mid-Upper Cretaceous Phase (Santa Lucian Orogeny)

The Upper Cretaceous strata in the Coast Ranges are divisible into two groups, the Pacheco and the Asuncion, which together make up the Chico (Taliaferro, 1943b). See Fig. 17.2. The Pacheco consists in

Fig. 17.12. Evolution of structure along cross section through south central part of Adelaida quadrangle, Calif., showing relations between various units of Cretaceous and relations to older and younger rocks.
1. Structure as it exists at present.
2. Structure along same section during deposition of Middle Miocene.
3. During deposition of Asuncion, late Upper Cretaceous.
4. During deposition of Jack Creek formation, early Upper Cretaceous.
(After Taliaferro, 1944.)



the central Coast Ranges of 7000 to 8000 feet of gray sandy shales, silts, sandstones, and conglomerates. If it was not removed by erosion before the Asuncion group was deposited, it rests on the erosion surface that followed the Diablan orogeny. The Pacheco sediments may be less widely distributed in the central Coast Ranges than the Asuncion but probably more widely in the northern Coast Ranges (Taliaferro, 1943b).

The Pacheco and Asuncion groups are separated by an unconformity which in places in the central Coast Ranges is as angular as 80 degrees. See Fig. 17.12. The disturbance represented by this unconformity has been named the Santa Lucian by Taliaferro (1943b). Where the Asuncion laps over older rocks than the Pacheco, which it does in a number of places, it is difficult if not impossible to distinguish the two disturbances —in fact, to recognize that more than one disturbance has occurred (Taliaferro, 1943b).

The Santa Lucian orogeny was strongest in the Santa Lucia Range and died out eastward. During the orogeny, the Gabilan mesa rose for the first time (Taliaferro, 1944). This has been called the Diablo uplift by Reed (1933). Another land projection into the general north-south trough lay to the south and has been called Catalina. At the north various authors have recognized the Klamath Island or Klamathonia. All three are here treated as peninsulas, branching off the volcanic archipelago, which as a whole has been called Pacifica. See the tectonic map of the Late Cretaceous, Plate 12.

As with other diastrophisms in California, the Santa Lucian appears to have taken but a relatively short time. Although there was deep erosion and widespread stripping, subsidence again took place, and the sea spread rather rapidly over an area of considerable relief. The latest Upper Cretaceous, the Asuncion, is the most widespread Cretaceous unit in the Coast Ranges. The Asuncion is predominantly coarse grained, being made up of arkosic sandstone and coarse conglomerates perhaps 10,000 feet thick. Fine sediments increase eastward. Franciscan debris increases toward the west. Near the present coast, the basal conglomerates contain large angular to subrounded blocks of Franciscan chert, basalt, diabase, and sandstone, as well as well-rounded pebbles, cobbles, and boulders of the ancient crystalline complex (Sur series and Santa Lucia granodiorite).

To the east in what is now the great valley of California, the Upper Cretaceous deposits have been divided into twelve foraminiferal zones, and these grouped into seven stages (Goudkoff, 1945). During the first three stages, the sea was transgressive eastward on the early Sierra Nevadan landmass, and reached a maximum distance at the end of the third stage except in the most northerly part. Near the end of the Upper Cretaceous (beginning of seventh stage) a low land barrier just south of Stockton divided the region into two basins. The extent of the barrier westward into the site of deposition of the Chico strata has not been worked out. During the earlier stages, the sediments came from the west as pointed out by Taliaferro, but in the later stages considerable material came from the east according to Goudkoff, and some from the northwest. The eastern source suggests slight uplift in the closing phase of the Cretaceous in the Sierra Nevadan landmass.

Evidence of igneous activity is present in many formations of the Mesozoic and Cenozoic in the central Coast Ranges of California, and Taliaferro, emphasizes the fact that volcanism was nearly continuous in one place or another nearby during these eras.

COLUMBIA SYSTEM

Extent

The term Columbia system will here be used to signify the mountains and troughs of the Mesozoic era in British Columbia, southeastern Alaska, the Yukon, Washington, western Idaho, and eastern Oregon. It is defined approximately by the extent of the Triassic and Jurassic troughs and the volcanic archipelago on the west that supplied much of the material to the troughs. In many respects it is a parallel, if not a continuation, of the great Sierra Nevada and Ancestral Coast Range systems of the United States. See tectonic maps, Plates 10, 11, and 12, and Fig. 17.13.

Triassic and Early Jurassic Phase

In the southern interior of British Columbia and more particularly southward from Kamloops Lake, strata, presumably of Triassic age, are widely displayed. See map, Fig. 17.14. This assemblage is generally re-

ferred to as the Nicola series and consists largely of volcanic intrusives and effusives, tuffs, and agglomerates with argillites and limestone at several horizons. The total thickness in places is 10,000 to 15,000 feet, but the uppermost part may be of early Jurassic age. In the vicinity of Kamloops, the Triassic beds with a basal conglomerate rest without angular unconformity on Carboniferous beds, and possibly the same general relation holds elsewhere. The Triassic strata occur also west of the Fraser River, and consist of dark green massive andesite and basalt, chert, argillite, limestone, and tuffaceous shales (Cairnes, 1936). Volcanic rocks with minor amounts of sediments occupy a great part of Vancouver Island and of Queen Charlotte Islands. See map, Fig. 17.18. The bedded character of the fragmental volcanic rocks, the presence of a few limestone members which in some places are several thousand feet thick, and the marine fossils found in tuffs and other sediments indicate that the general assemblage is marine in origin. It seems probable that the beds formed in a sea which, like that of Carboniferous time, extended over the greater part, if not all, of the Canadian Cordilleran region.

In Queen Charlotte Islands, a thick clastic series with some pyroclastic material ranges in age from Late Triassic to Jurassic and grades upwards into a volcanic assemblage of tuffs and effusives, 5000 feet or more thick. The tuffs are fossiliferous, evidently were laid down in the sea, and are of Jurassic and, presumably, Middle Jurassic age. The thick assemblage of Triassic volcanics and sediments which is widely displayed over Vancouver Island and known as the Vancouver group also may be succeeded by beds of Jurassic age. Strata, resembling the Vancouver group and related series but in places much metamorphosed, occur at intervals along the mainland coast and as included masses in the granitic rocks of the Coast Range. Along the eastern side of the Coast Range, within the basins of Nass and Skeena rivers, is a thick assemblage of sedimentary and volcanic rocks known as the Hazelton group and which, as indicated by an imperfectly known flora and fauna, is of Jurassic, possibly Mid-Jurassic, age. The proportions of sedimentary and volcanic material composing the Hazelton group varies from district to district with some indication that the nonvolcanic sedimentary rocks become more and more preponderant as the formation is followed eastward from the margin of the Coast

Fig. 17.13. Batholiths of the North American Cordillera. Heavy dashed lines indicate axes of anticlinoria, synclinoria, belts, and general trends of the late Jurassic and early Cretaceous phase. The dashed line north of the Nelson batholith is an anticlinorium in Proterozoic strata and may be Laramide in age.

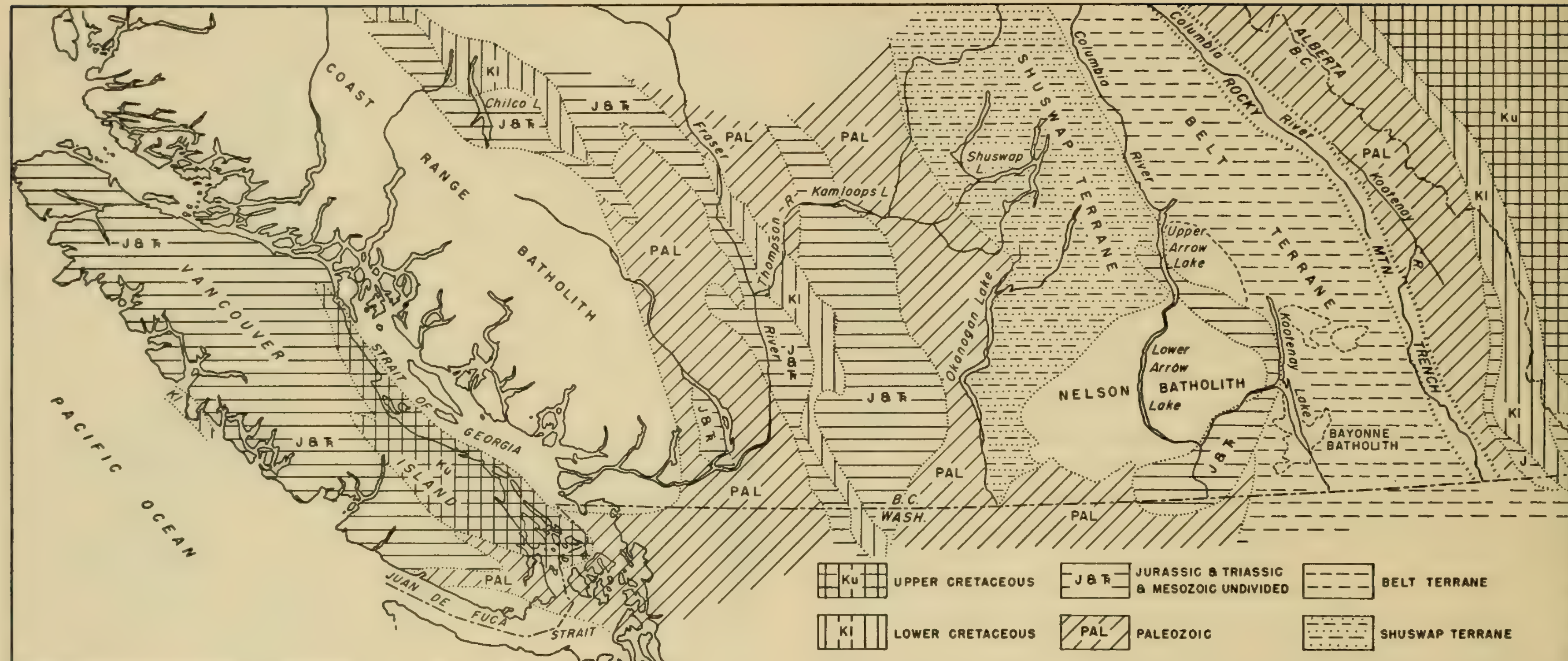


Fig. 17.14. Generalized distribution of major rock units of southern British Columbia. After Smith and Stevenson, 1955. The Shuswap terrane has some Paleozoic and Belt outcrops and the Belt terrane has some small patches of Paleozoic. All are invaded by great plutons, not shown, except the Upper Cretaceous of Vancouver Island and the strata east of the Rocky Mountain trench.

Range. The strata are, in part at least, marine; but the presence of plant remains and other features indicates that the sea was shallow and that possibly some parts of the assemblage may be nonmarine. In southern British Columbia, various assemblages of sediments or of sedimentary and volcanic strata are known, and others are believed to be of Jurassic age, as in the Kootenay Lake district where fossiliferous Jurassic beds rest on Paleozoic strata. In general, the Jurassic beds appear to be as widely, or even more widely, distributed than the Triassic; and as yet there is no evidence of any interval of orogenic movements that separated the two periods. Thick assemblages of sediments and pyroclastics, as well as great volumes of extrusive and intrusive volcanic strata, occur in northern British Columbia and southern Yukon and apparently correspond to the Hazelton group in the south. For more detail, see *Canadian Geological Survey, Economic Geology Series, No. 1,* 1957.

Daly (1912) describes Triassic strata in northwestern Washington that have a thickness between 3000 and 7000 feet. They are principally dark gray to black argillite, in part bituminous, generally associated with bands of gray to greenish gray sandstone and grit and in a few places with fine

conglomerate. The gritty beds are charged commonly with small angular fragments of black argillite. All the coarser types are decidedly feldspathic. Some of these sediments could probably be called graywacke.

In southeastern Alaska, Buddington and Chapin (1929) have noted numerous outcrops of Triassic rocks and others that may be Triassic. All the strata that carry fossils are Upper Triassic, and they seem to be divisible into three units, one consisting of sediments and the other two of volcanic rocks with a little intercalated sedimentary material. The volcanic formations are differentiated from volcanic formations of other periods on the basis of their faunas. Their character and structural relations over wide areas are insufficiently known, and their lithology is too similar, to separate them otherwise. They comprise green andesitic flows, breccia, and tuffs. The lava predominantly shows pillow structure but is in part amygdaloidal and in part polygonally jointed. Much of the breccia has a limestone matrix and is in considerable part the result of primary disaggregation of the radial-jointed pillows. The basal part of the volcanic rocks on Kuiu Island consists of interbedded limestone and green andesitic tuff and lava with local conglomeratic beds. On Kupreanof Island, the volcanic formation has a bed of conglomerate 150 to 200 feet thick in local areas at its base. The basal Triassic conglomerates and the unconformable relations to the Paleozoics have been discussed previously in the section on the late Permian or early Triassic orogenic phase.

On Kupreanof Island and the islands southeast of Kake, Upper Triassic sediments overlie the upper limestone division of the Permian and are overlain in apparent conformity by volcanic rocks of late Triassic age. Locally there is a thick bed of coarse conglomerate of the Upper Triassic volcanic rocks. On the northeast side of Kuiu Island, however, the Upper Triassic volcanic rocks overlie the lower division of the Permian without any apparent angular unconformity. The volcanic rocks of Kuiu Island also carry a different fauna from those on Kupreanof Island. Unconformities are indicated, therefore, not only at the base of the Upper Triassic, but within it (Buddington and Chapin, 1929).

The Triassic occurrences in British Columbia and Washington are so little known that unconformities within the beds assigned to this period, if they exist, are not known. It is of interest, however, to recall the unconformities below and above the Upper Triassic beds of western Nevada, and to note the same position of breaks in southeastern Alaska.

Another series of beds that was intruded by the great Coast Range batholith in southeastern Alaska has been assigned questionably to the Jurassic. Some of these beds may be Lower Cretaceous and some Triassic and Paleozoic. They have been divided into two groups for mapping purposes, namely, a predominantly sedimentary facies consisting of graywacke, black slate, and conglomerate; and a predominantly volcanic facies consisting of schistose greenstone made up of breccia, flows and tuffs, and black slate and graywacke.

These questionable Jurassic and Lower Cretaceous rocks are believed to overlie the Paleozoic and Triassic formations unconformably. A pronounced angular unconformity separates Jurassic from Devonian formations at the north end of Kupreanof Island, but where the Jurassic rests on the Triassic the break is more in the nature of a disconformity (Buddington and Chapin, 1929).

The Jurassic (or Lower Cretaceous) slate and graywacke appear to be much less metamorphosed than the older Mesozoic formations, but this is certainly in part due to their character. In all the formations, the argillaceous beds are most resistant to recrystallization, and their abundance in this series gives the Jurassic formations a misleading appearance of minor metamorphism. The pebbles and cobbles of the intercalated conglomerates are very markedly flattened as a result of very strong pressure.

To summarize, map, Fig. 17.14 may be referred to again. Triassic and Jurassic strata were deposited east of the present Canadian Rockies, and this basin of deposition was separated from a broad region of deposition by a Mesozoic geanticline which now is displayed chiefly as the Beltian terrane. The sediments of the eastern basin are miogeosynclinal and shelf types, whereas the sediments west of the geanticline are eugeosynclinal, and may have accumulated in several deep troughs. Due to the great batholiths that occupy much of the region of southern British Columbia it is apparently impossible to recognize the original extent of the basins or their number. Triassic and Jurassic strata occur in places from Kootenay lake westward to the Pacific Ocean. Since they are laden with volcanic

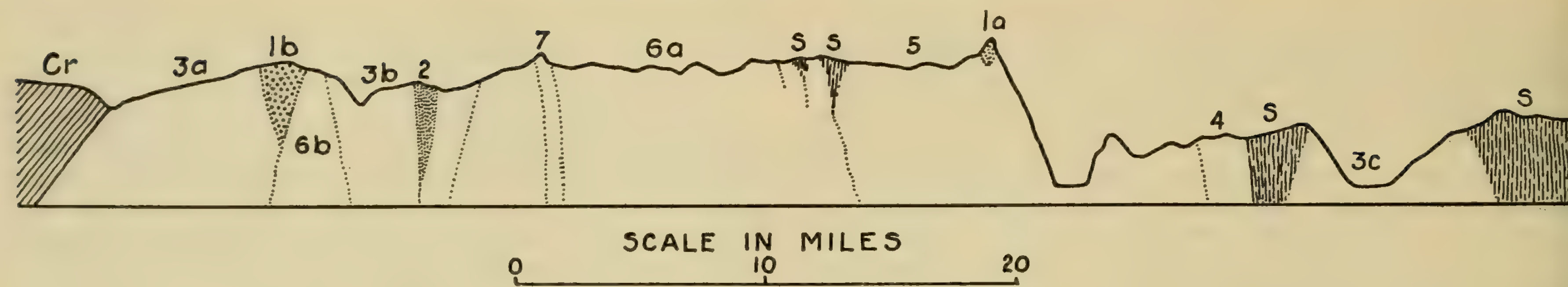


Fig. 17.15. Diagrammatic east-west section through the Okanogan composite batholith. s, schists and associated Paleozoic rocks; Cr, Pasayton lower Cretaceous arkose sandstones; Ia, Chopaka peridotite; Ib, Basic complex; 2, Ashnola gabbro; 3a, Remmel batholith, western phase; 3b, Remmel batholith, eastern phase; 3c, Osoyoos batholith; 4, Kruger alkaline body; 5, Similkameen batholith; 6a, Cathedral batholith, older phase; 6b, Park granite stock; 7, Cathedral batholith, younger phase.

The components of the batholith are numbered in order of intrusion. Vertical scale is exaggerated twice the horizontal. After Daly, 1912.

materials a west-lying volcanic archipelago and orogenic belt may be postulated similar to that of California and western Nevada, but possibly the regions of sedimentation were complicated by geanticlines which also were sites of volcanic activity.

Cretaceous strata have a more restricted distribution. Lower Cretaceous deposits of interior basin type occur in the Rockies and eastward. A narrow trough of them is recognized west and south of Kamloops Lake in mid-interior and east of the great Coast Range batholith. A small deposit occurs on the west coast of Vancouver Island. It would appear that the Belt geanticline had widened westward from Jurassic time, and perhaps another broad geanticline existed in the site of the Coast Range batholith and Vancouver Island.

In Late Cretaceous time the entire interior from the Rocky Mountains to the Strait of Georgia had become emergent and only flanking deposits accumulated. This was undoubtedly a consequence of the great batholithic intrusions and previous compressional orogeny of the broad cordilleran region.

Nevadan Orogeny

Batholiths of the International Border. An almost continuous succession of batholiths stretches more than 350 miles along the international border between Washington and British Columbia. These have been described by Daly (1912), Smith and Calkins (1904, 1906), and Smith (1904), and later studies have been made by Waters, Krauskopf, Campbell, and Pardee (see references in Waters and Krauskopf, 1941). It is highly probable that the plutons form the basement southward under vast areas of lavas of the Columbia Plateau because granitic rocks appear in the Ochoco–Blue Mountains uplift of central and eastern Oregon midway to the Klamaths and Sierra Nevada (Waters, 1933) and also southeastward under more lavas through the Thatuna batholith to the great Idaho batholith.

Intrusions of peridotite, gabbro, and diorite are associated with the prevailing granodiorite of the great batholiths. Quartz monzonite, quartz diorite, and granite are locally widespread. In certain areas where the succession of intrusions has been worked out, it is a cycle similar to that of the Sierra Nevada, viz., first the smaller bodies of ultrabasic rock, then gabbros and diorites, and finally the great granitoid bodies. Pegmatites are rare in the batholiths along the border in Washington, but aplite masses locally of almost batholithic proportions crosscut the earlier intrusions. The borders of the batholiths commonly show discordant relations to the country rock, and extensive masses of contact breccia are found along the intrusive margins (Waters, 1933). Some of the best examples of dis-

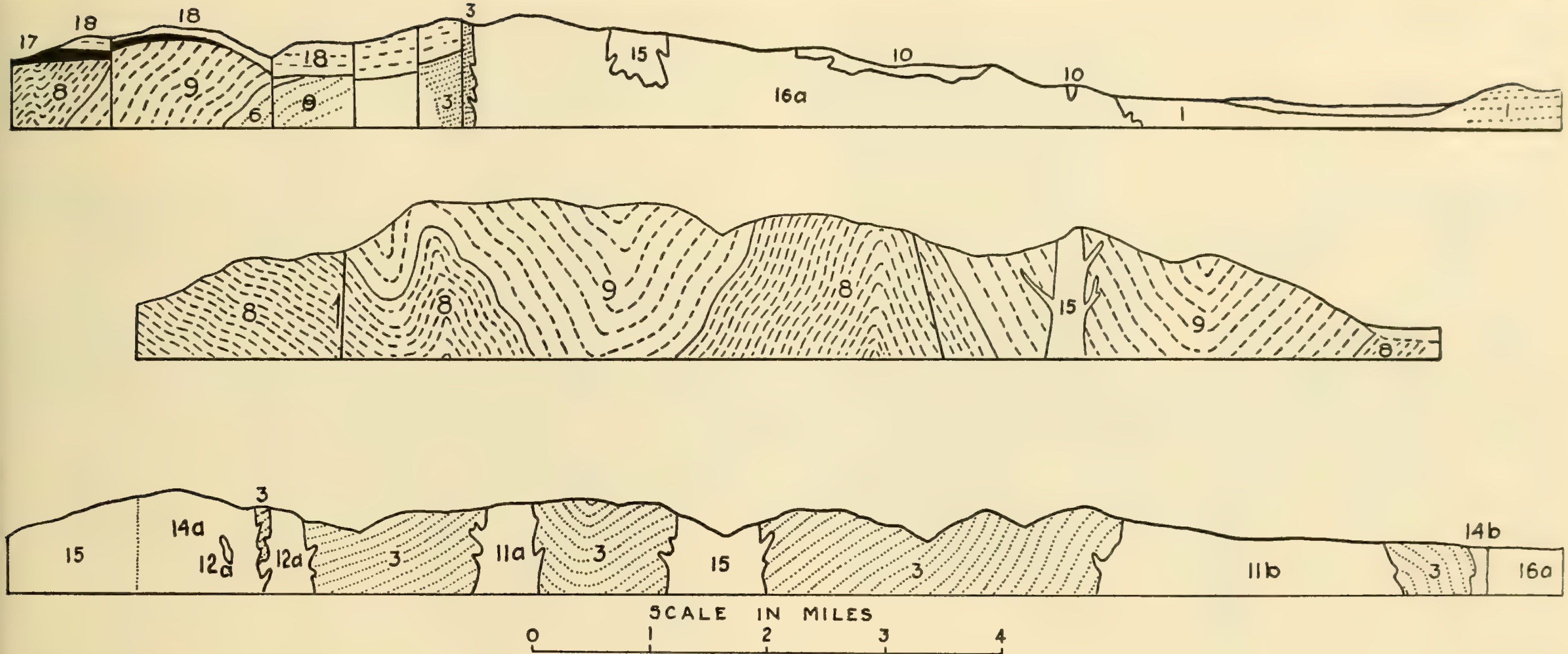


Fig. 17.16. Cross sections in the Similkameen District, British Columbia, Latitude 49, Longitude 120. 1, Vaseaux fm., paragneiss, schist, quartzite; 3, Koban group, schist, greenstone; 6, Barslow fm., argillite; 8, Shoemaker fm., chert, some tuff, greenstone; 9, Old Tom fm., greenstone, basalt flows, sills, bosses, some diorite; 10, altered rocks of dioritic composition; 11a, Osoyoos granodiorite; 11b, Fairview granodiorite; 12a, hornblendite; 12b, pyroxenite; 14a, Kruger syenite; 14b, Oliver syenite; 15, granodiorite; 16a, Oliver granite; 17, Springbrook fm., conglomerate, some sandstone and shale; 18, maroon fm., basaltic lava, some breccia, tuff, conglomerate.

1 and 3, Carboniferous. 6, 8, and 9, Triassic or older. 11 and 12, Jurassic (?). 14, 15, and 16, Jurassic and (or) younger. 17 and 18, Eocene. After Daly, 1912.

cordant batholiths, as well as some of the most conclusive evidence of stoping, may be found in the northern Cascades (Daly, 1912). See Figs. 17.15 and 17.16.

The igneous bodies generally designated by the names Osoyoos and Colville batholiths (see map, Fig. 17.13) are really complex associations of eight plutons. Contact metamorphism is intense near some but almost absent near others (Krauskopf, 1941). Detail along the border of the Colville batholith has been worked out by Waters and Krauskopf (1941). See Fig. 17.17. The batholith is a complex plutonic mass that intrudes folded and dynamometamorphosed sedimentary and volcanic rocks of late Paleozoic and Triassic age. Along the sharply discordant contact, the wall rocks are much fractured and granulated, but contact metamorphism is slight or absent. The batholith is remarkably heterogeneous, both structurally and petrographically. A central mass of structureless granodiorite grades outward into a belt of foliated igneous rock which commonly shows intricate swirling of the foliation. The swirled rocks grade into a peripheral belt of

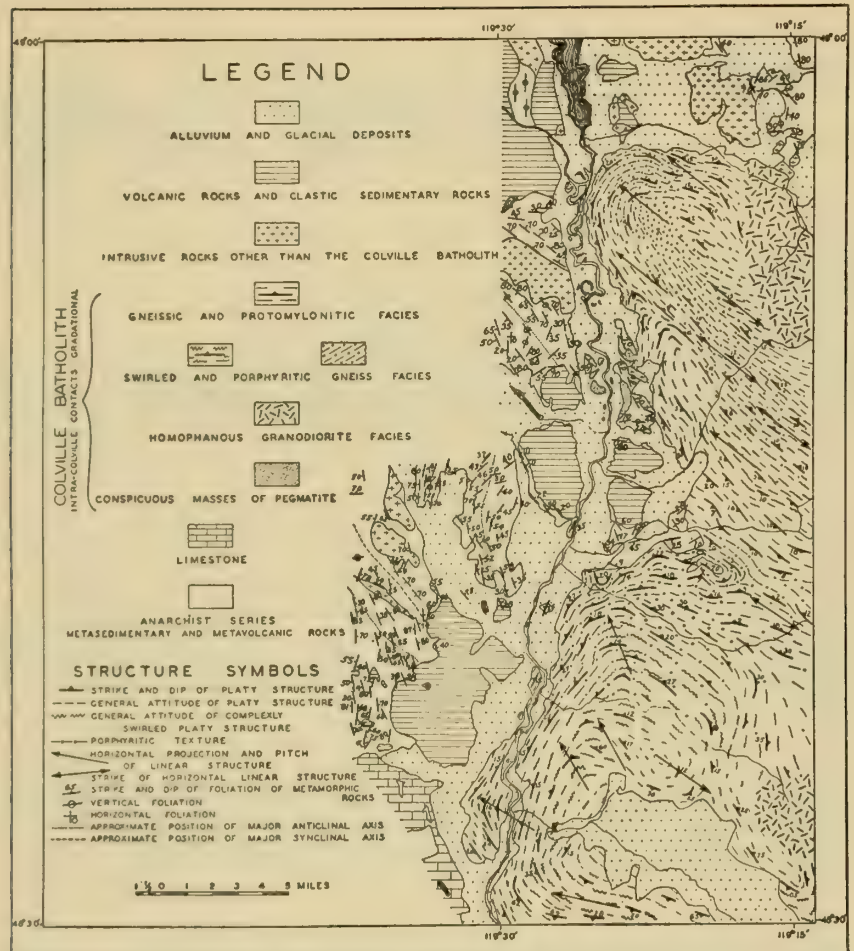


Fig. 17.17. Structural map of part of the border of the Colville batholith. Reproduced from Waters and Krauskopf, 1941.

variable but well-foliated migmatitic gneisses (magmatic injection and replacement) characterized by severe granulation of the constituent minerals. Over broad zones this rock is a mylonite (crushed and rolled-out streaky powder); locally recrystallization has produced types resembling metamorphic granulites. That the crushing was protoclastic (localized along the contact) and not due to regional metamorphism following the solidification of the batholith, is proved by the relations with the wall rocks and by the widespread cementation of the broken materials by films and stringers of undeformed quartz and microcline.

Along the contact between the approximately contemporaneous Osoyoos and Colville batholiths occurs a narrow belt of heterogeneous syenite with highly complicated internal structure. This is believed to be a hybrid rock formed by the action of magmas and emanations from both batholiths upon a thin septum of wall rock.

The Coast Range Batholith and Related Structures. The Coast Range batholith extends for more than 1100 miles from Fraser River in British Columbia northwestward into Yukon Territory. See Figs. 17.13 and 17.18. From Vancouver to Skagway on the mainland, the batholith forms the backbone of the Coast Range and is exposed either at the shore line or a short distance inland. Outlying dikes, stocks, and batholiths believed to be of the same general period of intrusion as the main batholith and genetically allied to it are found locally on Vancouver Island and the Queen Charlotte Islands, and abundantly throughout most of the Alexander Archipelago. The Coast Range batholith is the largest on the North American continent, aside possibly from certain ones of Precambrian age.

It is widest south of Skeena River in British Columbia, where it reaches 110 miles east and west. In southeastern Alaska, it is 35 to 60 miles wide.

Buddington (Buddington and Chapin, 1929) discusses the batholith in southeastern Alaska in respect to six approximately parallel belts, namely, the border zone east of the Coast Range batholith, the Coast Range batholith of the mainland, the Wrangell-Revillagigedo metamorphic belt, the Prince of Wales–Chicagof belt, the Kuiu-Heceta belt, and the Dall-Baranof belt. See map, Fig. 17.19.

The eastern border zone is conspicuous for its absence of contact metamorphism on a regional scale. Even local metamorphism is meager. The border zone rocks are closely folded; argillaceous rocks have been changed to slaty types, and locally andesitic volcanic rocks to greenstone; but there is practically no phyllite and no crystalline schist away from the immediate contact of the intrusive bodies.

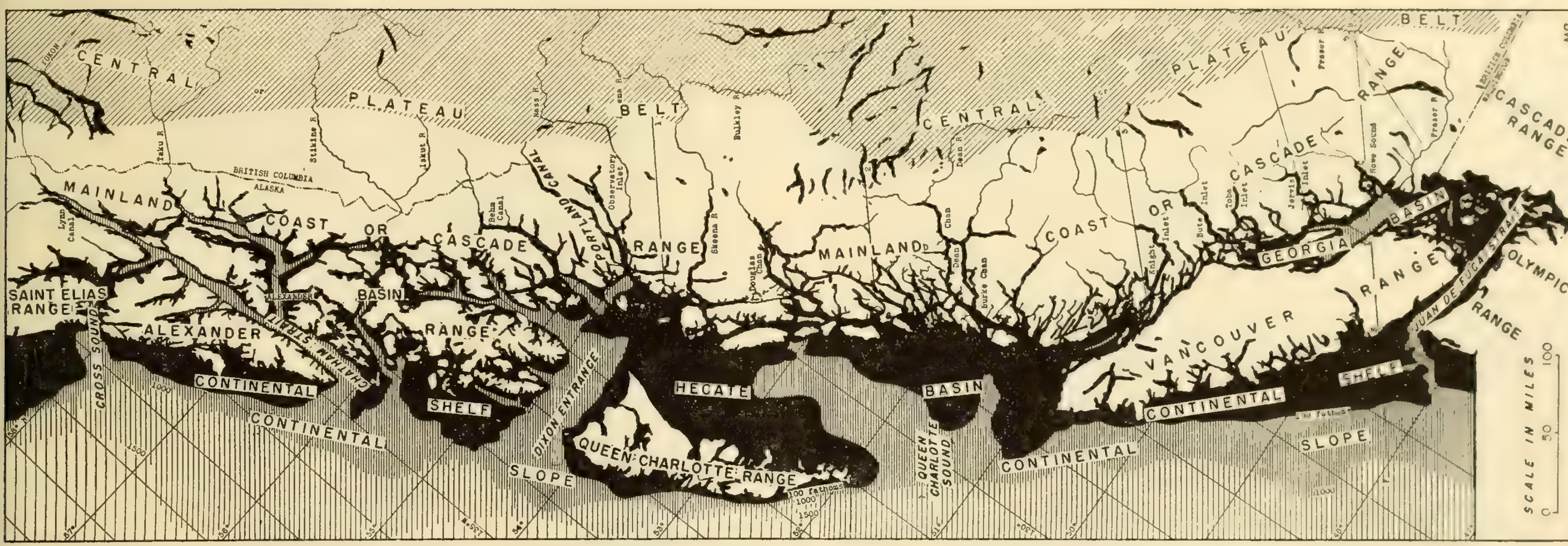


Fig. 17.18. Geomorphic divisions of British Columbia and southeastern Alaska showing the divisions of the western, or Pacific mountain, belt of the Cordilleran region. To bring the fiord system into prominence the sea within the 100-fathom line is shown in solid black. Within the black area are many basins which are deeper than 100 fathoms, especially in the fiord channels. After Peacock, 1935.

Some stocks or batholiths within the eastern border zone are imperfectly known. Between the Skeena and Nass rivers the border of the main batholith is irregular with apophyses and outlying stocks of granodiorite. Between Nass River and the Portland Canal, the border is fairly straight. In the Hyder district, a mass of hornblende granodiorite has been called the Texas batholith (Buddington and Chapin, 1929). It is probably adjacent to the main Coast Range batholith and is cut by many dikes of younger quartz monzonite and granodiorite. It is locally intensely crushed, evidently from the thrust of the younger intruding magma. Although definitely older than the intrusions that cut it, the Texas batholith is probably post-Jurassic, because it is similar to a nearby quartz diorite at the head of Hastings Arm and Observatory Inlet which is intrusive into the "Bear River series" and Nass argillite of Jurassic age.

Intrusive bodies are also known in the Atlin and Whitehorse districts, but for the most part the vast region northwest of the Hyder district is unknown.

The great Coast Range batholith itself is inadequately known, but Buddington's description (Buddington and Chapin, 1929) for the section between the Portland Canal and the Stikine River is illuminating. The southwest border facies in a belt 5 to 15 miles wide, has the average composition of a granodiorite, and is composed predominantly of granodiorite, quartz monzonite, and quartz diorite; the eastern border facies, 10 to 15 miles wide, is quartz monzonite. Dolmage (1923) reports that the more silicic variations lie in the center of the batholith south of Portland Canal as Buddington finds to the north, but that there are exceptions. The changes from one type of rock to another appear to take place rather abruptly, but no evidence of brecciation of one variant by another has been seen, except in the small masses of gabbroic and ultrabasic rocks.

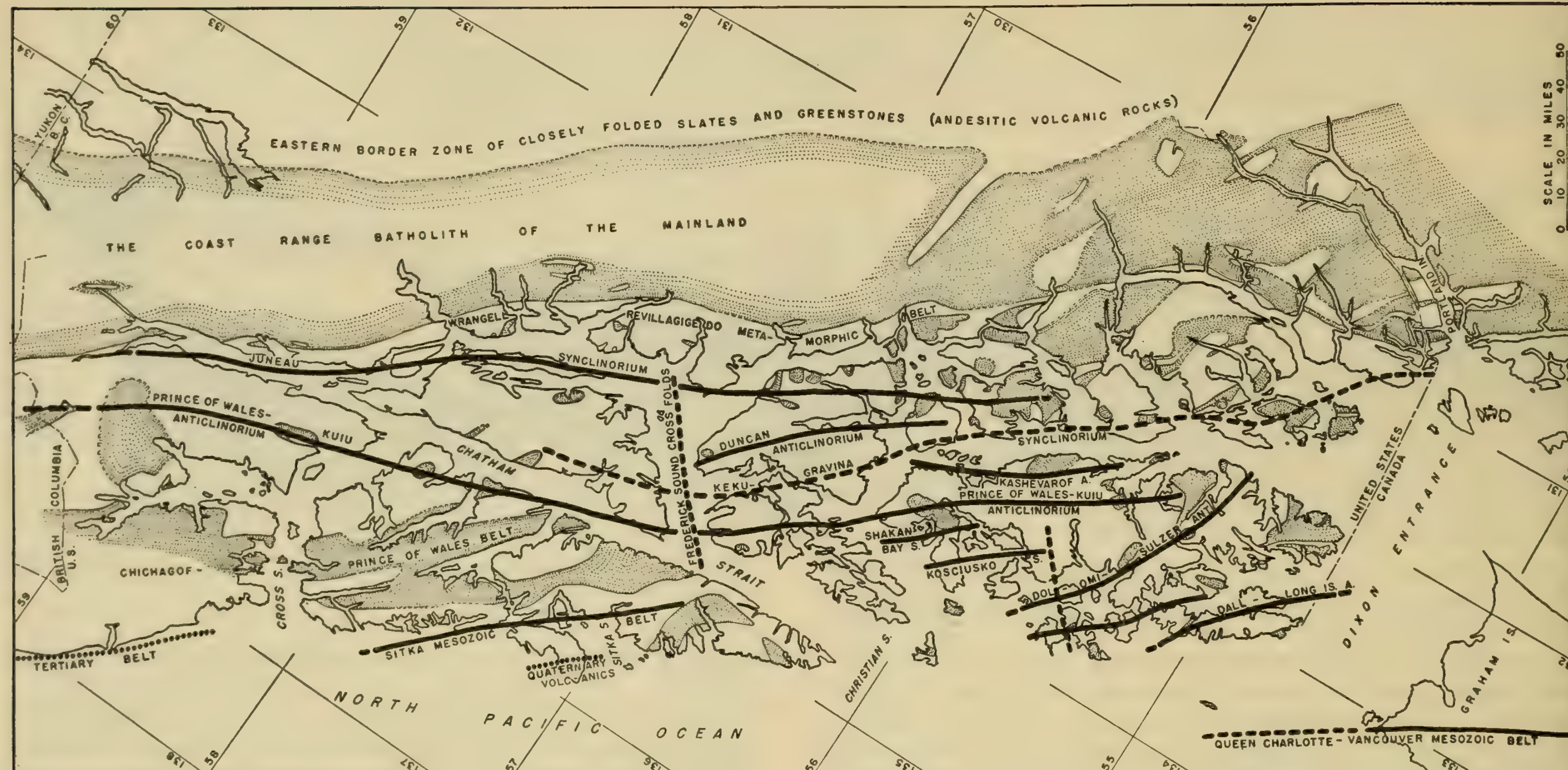


Fig. 17.19. Tectonic map of southeastern Alaska and adjacent parts of Canada. After Buddington and Chapin, 1929. Stippled areas are related bodies of the Coast Range batholith. The solid heavy lines are the axes of anticlinoria and synclinoria that were formed at approximately the same time that the batholith and its satellites were intruded. The dashed heavy lines are the axes of broad Tertiary arches and sags, and also in part of folds associated with the intrusions.

Buddington believes the variations are closely related to interlocking batholiths which together build what is known as the Coast Range batholith.

The great batholith is paralleled on the west by a belt of injection gneiss, crystalline schist, and phyllite intruded by abundant batholiths, stocks, sheets, and dikes, believed to be outlying masses genetically associated with the main batholith. See cross sections, Fig. 17.20 This zone as previously mentioned is the Wrangell-Revillagigedo belt of metamorphic rocks. It is narrow and loses its individuality at the north but widens and is very well defined toward the south. Near the mouth of Gastineau Channel and west of Thomas Bay this belt has a width of about 13 miles; opposite the mouth of Stikine River, about 25 miles; and at the south end of Revillagigedo Island, about 35 miles. This composite belt of sedimen-

tary and intrusive rocks appears to swing from a northwest strike northwest of the Cleveland peninsula to a north-south strike south of the peninsula. This is not due to a change from the prevailing northwest strike of the beds but to a farther west penetration of the intrusive masses at the south end. The differences between the smaller intrusive masses in the metamorphic belt and the quartz diorite of the western border of the batholith are slight.

West of the metamorphic belt is the Prince of Wales–Chicagof belt in which intrusive masses are common but less quartzose, and the country rock consists predominantly of slate, limestone, graywacke, greenstone, and dynamically metamorphosed schistose rocks with locally some crystalline schist and marble. The metamorphism is much less advanced than in the Wrangell–Revillagigedo belt, though locally adjacent to large igneous bodies it may be intense. The belt is 40 miles wide on the north, but not much more than 5 miles wide on Kupreanof Island; it widens on Etolin Island, and is about 25 miles wide through Prince of Wales Island. The intrusive rocks of this belt differ, in general, from those to the east in that they are predominantly diorite, rather than quartz diorite, and that differentiates of highly contrasted composition are more adundant.

The Kuiu–Heceta belt is next west and is characterized by the least metamorphism of any of the belts, by the fewest intrusives, and as a result, by the best preserved and oldest fossils in its strata. The belt includes the western fringe of the north half of Prince of Wales Island, the north end of Dall Island, San Fernando, Heceta, Tuxekan, Kosciusko, and Kuiu islands, Kupreanof Island with the exception of the Lindenberg peninsula, and the southwestern part of Admiralty Island.

The Dall–Baranof belt is the westernmost of the six belts of the great batholith with its satellites, and is characterized again by numerous stocks and batholiths. It includes Dall, Forrester, Suemez, Baker, Lulu, Noyes, Warren, Coronation, and Baranof islands. The intrusive rocks on the average are more silicic and carry less of ferromagnesian minerals than the average of the Prince of Wales–Chicagof belt. Quartz diorite and, to a lesser extent, granodiorite predominate.

In the main batholith, the rocks are prevailingly gneissoid. The banded character is most accentuated near the borders of the batholith or near inclusions within the batholiths. Local schistose zones are found along intensely sheared narrow bands. The gneissic structure is for the most part interpreted by Buddington as primary, but still the batholith was stressed considerably after its complete solidification. Yielding occurred by mashing along local belts or zones, which may be of considerable width and great length (75 miles or more), or by intensive shearing along narrow zones, or by slipping along many planes of various orientation throughout a zone. A belt of highly mashed rock 15 miles wide is crossed by Stikine River from the head of Little Canyon to and below Flood Glacier.

In some places west of the main batholith, extensive belts, including intrusive igneous stocks, dikes, and sills, constitute a local shear zone or zone of close folding; the larger masses of igneous rocks may show considerable mashing, and the thin sills may be closely folded together with the schists.

After reviewing pertinent studies on the succession of the intrusions that make up the great batholith all the way from Vancouver to Cross Sound, Buddington draws the following conclusions:

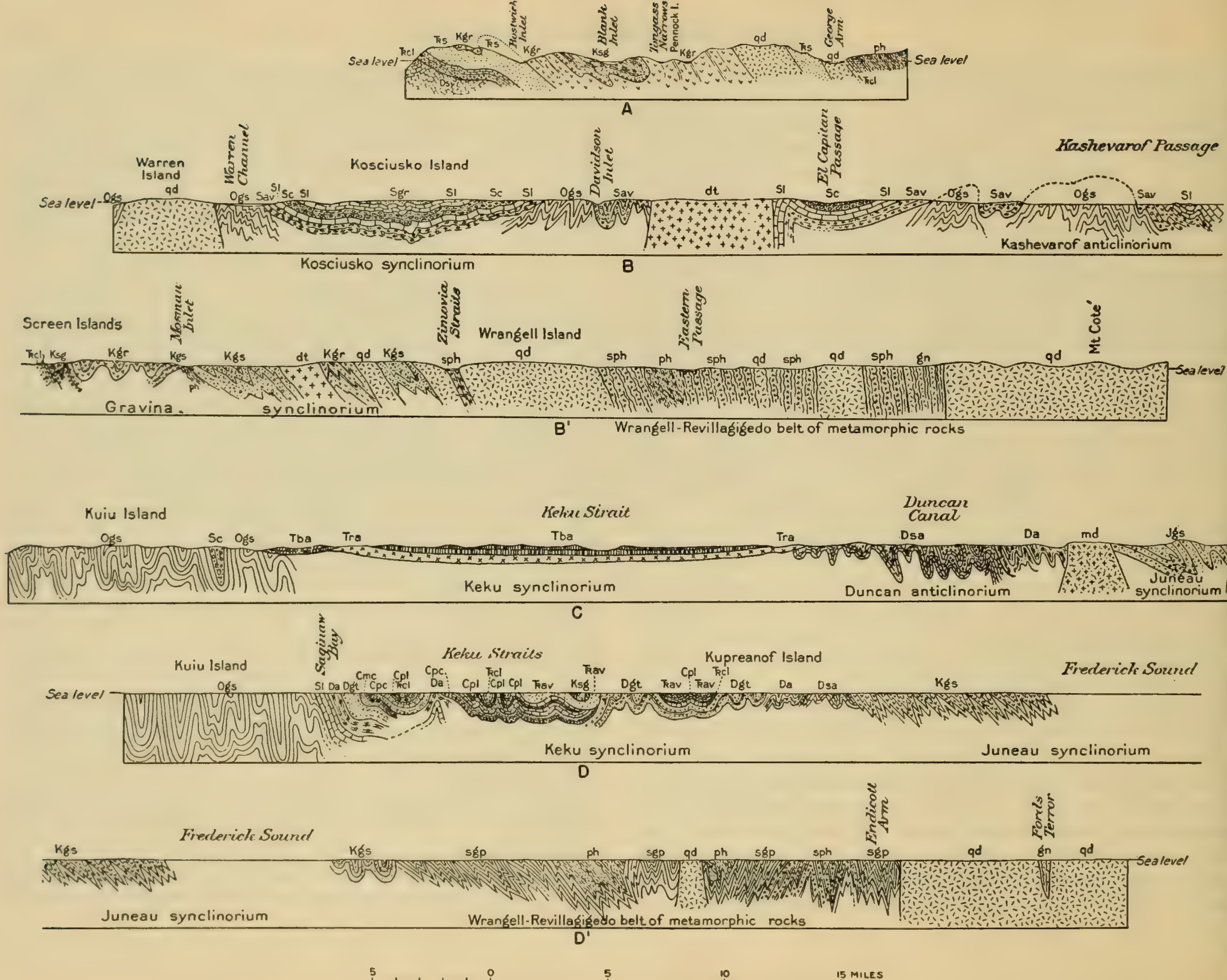
West of the main batholith a group of ultrabasic intrusive rocks is present in considerable volume. These include hornblendite, pyroxenite, dunite, peridotite, and intermediate variants; they are older than the more silicic-alkalic types.

Diorite and gabbro-diorite occur both as discrete stocks and batholiths, and also to a minor extent as marginal variations of quartz diorite and granodiorite. Locally gabbro-diorite and augite gabbro are the marginal phases of diorite. The gabbro-diorite and gabbro are locally intruded by diorite, quartz diorite, and more silicic-alkalic types.

Granodiorite in stocks and small batholiths may show marginal variants of quartz diorite, gabbro-diorite, diorite, monzodiorite, and very rarely of syenite. A decrease in potassic feldspar and quartz locally on the margins is a common feature.

Granite is the youngest of the major members of the plutonic complex, and is uniformly found with intrusive contacts through the older members.

Buddington also point out that west of the Sierra Nevada batholith in the Coast Ranges of California there is a similar older group of ultrabasic


Sea level
Bostwick Inlet
Blank Inlet
Tongass Narrows
Pennock I.
George Arm
A
Warren Island
Warren Channel
Kosciusko Island
Davidson Inlet
El Capitan Passage
Kashevarof Passage
Kosciusko synclinorium
Kashevarof anticlinorium
B
Screen Islands
Moira Inlet
Zimovia Strait
Wrangell Island
Eastern Passage
Mt Cote
Gravina synclinorium
Wrangell-Revillagigedo belt of metamorphic rocks
B'
Kuiu Island
Keku Strait
Duncan Canal
Keku synclinorium
Duncan anticlinorium
Juneau synclinorium
C
Saginaw Bay
Keku Straits
Kupreanof Island
Frederick Sound
Keku synclinorium
Juneau synclinorium
D
Endicott Arm
Fords Terror
Juneau synclinorium
Wrangell-Revillagigedo belt of metamorphic rocks
D'
5 0 5 10 15 MILES

intrusive rocks and a younger group of more silicic-alkalic intrusive rocks.

Structure of Southeastern Alaska. Most pervading of any structural feature is the isoclinal folding. In a belt 15 to 30 miles wide adjacent to the west side of the Coast Range batholith, the isoclinal folds are slightly or markedly overturned toward the southwest. See Fig. 17.20. In general the axes or axial planes of the isoclinal folds escape detection, and uniform dips occur through wide intervals. In the Wrangell–Revillagigedo belt of metamorphic rocks near the batholith, the dip of the schist and gneiss is 60 to 90 degrees northeast. In the outer part of the belt, the dip is 30 to 50 degrees northeast. On the east side and at the northeast end of Prince of Wales Island, the isoclinal folds are overturned toward the northeast (Buddington and Chapin, 1929).

Out of the numerous isoclinal and close folds, a number of anticlinoria and synclinoria may be recognized. Two dominate the structure of the Alexander peninsula, namely: the Junean synclinorium next west to the great batholith, and west of the synclinorium the Prince of Wales–Kuiu anticlinorium. Both synclinorium and anticlinorium seem to be divided into branching or parallel synclinoria. The axes of all the major fold complexes are shown on the map of Fig. 17.19.

The formations exposed in the Prince of Wales–Kuiu anticlinorium are almost exclusively Paleozoic, and make up a belt 40 to 50 miles wide. In the trough of the synclinorium the Upper Jurassic and Lower Cretaceous formations are exposed. The Keku–Gravina synclinorium is dominantly a shallow downwarp of Tertiary formations. They rest unconformably on Mesozoic formations which in general are folded into the great Junean synclinorium.

On the west flank of the Prince of Wales–Kuiu anticlinorium in the Sitka district is the Sitka Mesozoic belt. A western belt of Tertiary sediments and volcanics is exposed along the shore north of Cross Sound, and the southern half of Kruzof Island on the north side of Sitka Sound is composed of Quaternary volcanics (Mt. Edgecombe).

Through Frederick Sound is an axis of cross folding. Jurassic beds are exposed at intervals along the south side of Admiralty Island and form a considerable part of the coastline. They appear to have been folded along two axes—the usual north-northwest axis and another about parallel to Frederick Sound—that is, east-west. The gently dipping Tertiary lava beds about Frederick Sound seem to represent a broadly folded anticline from the center of which they have been eroded away. The axis of the anticline strikes northeast, approximately in the direction of the Mesozoic cross fold, and Buddington suggests that the forces in Tertiary time were oriented almost at right angles to those that effected the folding in pre-Tertiary time.

Structure of the Island Ranges of British Columbia. For a clear discussion of the topographic elements of British Columbia refer to Peacock (1935). Under the present heading, the area west of the Coast Range batholith in British Columbia is signified. No summary treatment of the folds and faults of this great island region has been written such as Buddington and Chapin's account of southeastern Alaska, although numerous reports of specific areas are available. They are chiefly *Summary Reports of the Geological Survey of Canada.* Even if possible, it does not seem feasible for the present writer to attempt a synthesis, but in general it appears that the same type of structure as in southeastern Alaska con-

Fig. 17.20. Structure section in southeastern Alaska, after Buddington and Chapin, 1929. A, across Gravina and part of Revillagigedo Islands, showing Triassic beds thrust over Devonian. B and B′, continuous section from Iphigenia Bay to the mainland. C, across Kuin and Kupreanof Islands to the mainland. D and D′, continuous section along the south side of Frederick sound to the mainland. Upper Jurassic of Lower Cretaceous intrusives: dt, diorite; md, monzodiorite; qd, quartz diorite. Metamorphic rocks, probably Ordovician to Jurassic or later, Wrangell-Revillagigedo belt: sgp, schistose greenstone and green phyllite; ph, phyllite; sph, md, crystalline schist and phyllite with beds of marble; gn, layered gneisses. Lower and Middle Ordovician: Ogs, indurated graywacke with slate, andesitic volcanics, chert, conglomerate, and limestone. Silurian: Sar, andesitic volcanics and conglomerate; Sl, limestone, with thick conglomerate, sandy beds or argillaceous beds (Sc); Sgr, predominant graywacke. Middle Devonian: Dsa, slate, limestone and chert with interbedded andesitic volcanics; Da, andesitic lava, breccia and conglomerate with limestone cobbles; Dgt, predominantly graywacke and tuffaceous beds; Dsr, sediments, including graywacke, conglomerate, slate, limestone and chert, with associated volcanics. Mississippian: Cmc, chert, quartzite and limestone. Permian: Cpc, conglomerate, limestone, sandstone, andesitic and basaltic volcanics; Cpl, limestone with white chert layers. Triassic: Ꞧcl, conglomerate, sandstone, and limestone; Ꞧs, slate with sandstone in upper part; Trav, andesitic volcanics, including breccias and lava flows locally interbedded with sediments. Jurassic or Cretaceous: Kgs, graywacke, slate, and conglomerate with tuff and limestone; Kgr, greenstone volcanics. Lower Cretaceous: Ksg, slate and graywacke with chert nodules, impure limestone, and conglomerate. Eocene: Tra, rhyolite and andesite volcanics, conglomerate, and dacite porphyry sills; Tba, basaltic and andesitic lava with some breccia and conglomerate. Quaternary: Qb, basalt and tuff.

tinues southeastward through the Island Ranges of British Columbia. The isoclinal folds, the prevailing northwestern strikes, and the steep northeastern dips are similar. The belt of metamorphic rocks that flanks the western margin of the batholith seems to continue a considerable distance southward into British Columbia. The synclinoria and anticlinoria do not seem to have been worked out, but perhaps the variety or number of such features is not exposed or not existent.

The Keku–Gravina synclinorium in Mesozoic rocks extends southeastward across the international border to Pitt Island, but seems to end in the great batholith before reaching Douglas Channel. The Dall–Long Island and Dolomi–Sulzer anticlinoria in the Paleozoic strata either die out southeastward or are covered by the waters of Hecate Strait, because on the west is a wide Mesozoic belt of Queen Charlotte Islands and ancouver Island. Lacking information, it can only be assumed that this belt is a broad synclinorium. It seems to correlate with the Sitka Mesozoic belt 200 miles to the northwest, but if so it must bulge westward around the Paleozoic anticlinoria of Dall and Prince of Wales islands.

There seems to be plenty of room for an anticlinorium and another synclinorium under Hecate Strait and Queen Charlotte Sound.

The Paleozoic belt striking nearly east-west on the southern end of Vancouver Island may mark another anticlinorium outside the Queen Charlotte–Vancouver Mesozoic belt.

Concordant Fracture System. The pattern of the great batholiths of the Upper Jurassic and Lower Cretaceous of western Mexico, the United States, Canada, and southeastern Alaska remind Peacock (1935) of the arc-and-cusp plan of the circum-Pacific orogenic belts. This feature has already been referred to. Since the batholithic rocks show little evidence of deformation, the curved plan appears to have originated during the emplacement of the igneous rocks and during the preceding orogenic events. The grain of the coastland, in British Columbia and southeastern Alaska, as defined by the folds and foliation, is longitudinal to the arcs, and therefore is intimately associated with the arcs in origin. Deformation during Cenozoic time has had little effect on the Mesozoic pattern.

Peacock (1935) recognizes the fiords and straight stretches of coastline to be the result of erosion controlled by a fracture system composed of two elements, viz., a concordant one and a discordant one in relation to the arc. The concordant system is composed of fractures parallel to the grain and normal to it, and the discordant of a north-south and east-west system. See Fig. 17.21 and compare with Fig. 17.18. The first was formed shortly after the solidification of the batholith; the second at the close of the Cretaceous. Dikes and mineralized veins follow the transverse fractures of the concordant system. Because of the fissure type of vein fill, the transverse fractures appear to be of tensional origin (Balk, 1937) and associated with the batholith. The main faults thus far recognized along which the fiords have been eroded are the Lynn Canal and Chatham Strait, but these apparently belong to the younger discordant system.

Peacock's (1935) analysis of the mechanics of the great fracture system is as follows:

> If the coastland be regarded as a tabular body of rigid material undergoing deformation by dominating horizontal forces acting from the northeast, causing differential horizontal displacement toward the southwest, with the development of an arc bent away from the dominant pressure, then tensile stresses, as in a bent beam, would develop in the advanced part of the arc. These stresses would be relieved by tension fractures running normal to the directions of maximum tensile stress and, therefore, transversely to the grain or radially to the arc. It is not to be expected that such tension fractures should follow strictly radial directions. Although generally transverse, such fractures might deviate considerably from directly transverse courses, because of irregularities in the mechanical strength of the region; they might also change abruptly from the transverse to the longitudinal direction, the weak direction of the grain, and thus develop a cranked course with rectangular elbows.
>
> With the mode of deformation suggested, shearing stresses would also be set up along vertical planes parallel to the grain, and these would be relieved by longitudinal shear fractures of the shear type.
>
> If formed in the manner outlined, the transverse fractures would be open fractures, and when mineralized they would appear as fissure veins. The longitudinal fractures would be closed fractures along which some horizontal differential movement would occur to relieve the shearing stresses. Mineralization of such ruptures would result in mineralized shear-zones such as Schofield has found usually to lie in the longitudinal direction. Both sets of fractures would provide ready-made planes of faulting when subsequent crustal unrest affected the region and caused differential movement between the already separated blocks.

It is also possible that thrust faults would form. If relief or elongation is easiest in the vertical direction, then shear planes would form which strike longitudinally and have dip-slip movement. Buddington mentions shear zones or faults with strikes of N 38° W. to N 60° W.

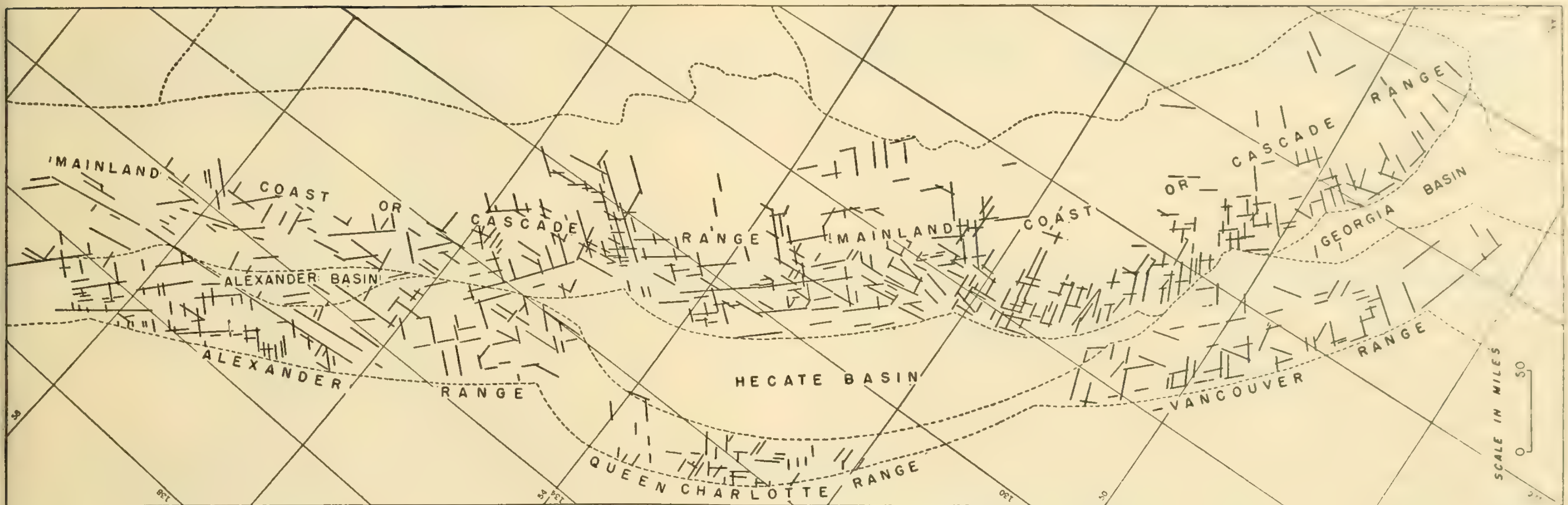


Fig. 17.21. Sculpture pattern of the coastland of British Columbia and southeastern Alaska obtained by drawing straight lines along all nearly straight fiord reaches, lake shores, stream courses, and portions of coastline. The complex pattern resolves itself into a concordant pattern consisting of lines running parallel and at right angles to the curving longitudinal grain, and a discordant pattern composed of lines lying north-south and east-west obliquely to the grain. All four directions are prominent directions of jointing. After Peacock, 1935.

Age of the Batholiths. The following summary of the age of the great batholiths is taken from Buddington and Chapin (1929).

The age of the Mesozoic intrusive rocks has not been definitely determined. To the northeast, on the east side of the batholith in the Whitehorse district, Yukon Territory, the intrusive rocks are reported by Cockfield to cut rocks of Middle Jurassic age and, therefore, to be probably of Upper Jurassic age or later. Hanson reports that on the east side of the batholith, in British Columbia, between Skeena River and Steward, the Coast Range batholith intrudes the Hazelton group (Jurassic) but does not intrude the Skeena (Lower Cretaceous) series. He says: "It is, therefore, probably mainly of Upper Jurassic age, but parts of the batholith may be of later age." Dolmage, in describing the Tatla–Bella Coola area, writes: "In Taseko Lake district, what appears to be the main Coast Range batholith cuts a thick series of coarse fragmental volcanic rocks in which the writer found plant remains, determined by E. W. Berry to be of Cretaceous age. . . . This evidence proves that this part at least of the batholith is younger than the lowest Cretaceous, and the evidence found in Tatlayoko Lake, Taseko Lake, and Bridge River districts strongly suggests that much of the eastern part of the batholith is of post-basal Lower Cretaceous." Cairnes suggests that at the southeastern part of the batholith, on the eastern border, there are intrusions of two ages. Masses of intrusive rocks that cut probable Jurassic beds are reported by him to be overlain unconformably by beds of Lower Cretaceous age, and the Lower Cretaceous beds are in turn cut by intrusions of pre-Tertiary age. On Vancouver Island the Mesozoic intrusive rocks are known definitely to be older than Upper Cretaceous.

In southeastern Alaska all the intrusive rocks classed as Mesozoic are definitely known to be older than the Eocene. On Chicagof Island intrusions of the Coast Range type are proved by Overbeck to cut fossiliferous beds of Upper Jurassic age. The writer is convinced that on Admiralty Island intrusions of the Coast Range type cut beds which, where not metamorphosed, carry the fossil *Aucella crassicollis* and which are therefore probably of Lower Cretaceous age. At the head of Portland Canal there is positive evidence of two epochs of intrusion; the older batholith cuts beds of the Hazelton series (Jurassic) and is in turn intruded by the quartz monzonite of the Coast Range batholith.

It is evident that for the most part the youngest beds with which the Mesozoic intrusive rocks are found in contact are of Middle or Upper Jurassic age; at a number of localities intrusive rocks of the Coast Range type cut Lower Cretaceous formations; there were at least two epochs of intrusion; and the Mesozoic intrusive rocks are all older than the Upper Cretaceous. So far as

southeastern Alaska is concerned, the writer is aware of no evidence to disprove the assumption that all the Mesozoic intrusive rocks may be of Lower Cretaceous age, but the data given for adjacent territory suggest that they may be in part of Upper Jurassic and in part of Lower Cretaceous age.

Follinsbee *et al.* (1957) report a potassium-argon age of the Coast Range batholith near Vancouver of 105 m.y. This would be at least midway up in the Lower Cretaceous, and corresponds well with the stratigraphic evidence above.

Relation of Batholiths to Folding. Regarding the relation of folding and intrusion, Buddington states:

> There is a most pronounced increase in the degree of crumpling, plication, foliation, and isoclinal folding as the border of the batholith is approached from the west, suggesting that the batholith has exerted a tremendous thrust. The manner in which the batholith has peeled off great slabs of schist constitutes further evidence. On the other hand, in the vicinity of the adjacent outlying stocks, sintering and compacting of the phyllite and slate as the contact is approached indicates that the cleavage and foliation are in part older than the intrusion.
>
> The data are inadequate for a solution of the problem. But if we assume that the intrusion of the batholith took place within the same general period as the Jurassic or Cretaceous folding, then it is probable that at least two factors were involved—an increased local intensity in the dynamic metamorphism above the location of the rising magma and a thrust exerted by the magma itself during its emplacement at horizons equivalent to those now exposed. Under the same stress and with other conditions the same, rocks will be much more highly deformed under higher temperature. Thus it might be that though stresses of essentially similar orders of magnitude affected beds both far to the east and far to the west of the present highly folded zones, the beds to the east and west, relatively much cooler, yielded by close folding and development of cleavage, whereas those in the intensely folded zone, at a higher temperature due to the rising magma with its advance wave of escaping highly heated vapors, yielded far more extensively. A preliminary foliated or cleaved character had thus already been induced before the arrival of the magma, which accentuated the dynamic effects by its own thrusting pressure and aided recrystallization by heat, vapors, and solutions.
>
> Another important factor appears to have been the structural relations which the invaded formations bore to the magma. For example, where they were in steeply dipping attitudes above the rising magma, conditions for penetration by magmatic solutions and vapors were favorable and metamorphism was correspondingly facilitated; such seems to have been the condition in the belt adjacent to the southern part of the batholith in southeastern Alaska. Where the contact plunges steeply the transfer of solutions and vapors was markedly

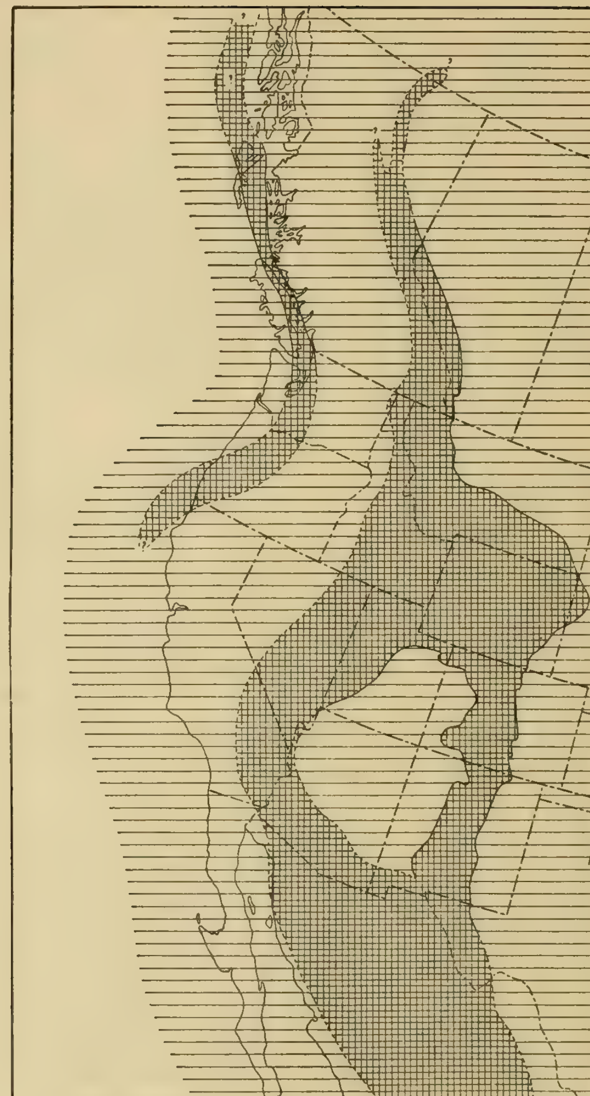


Fig. 17.22. Belts of the Laramide orogeny in the Rocky Mountains and the folded Upper Cretaceous trough of Oregon, Washington, and British Columbia.

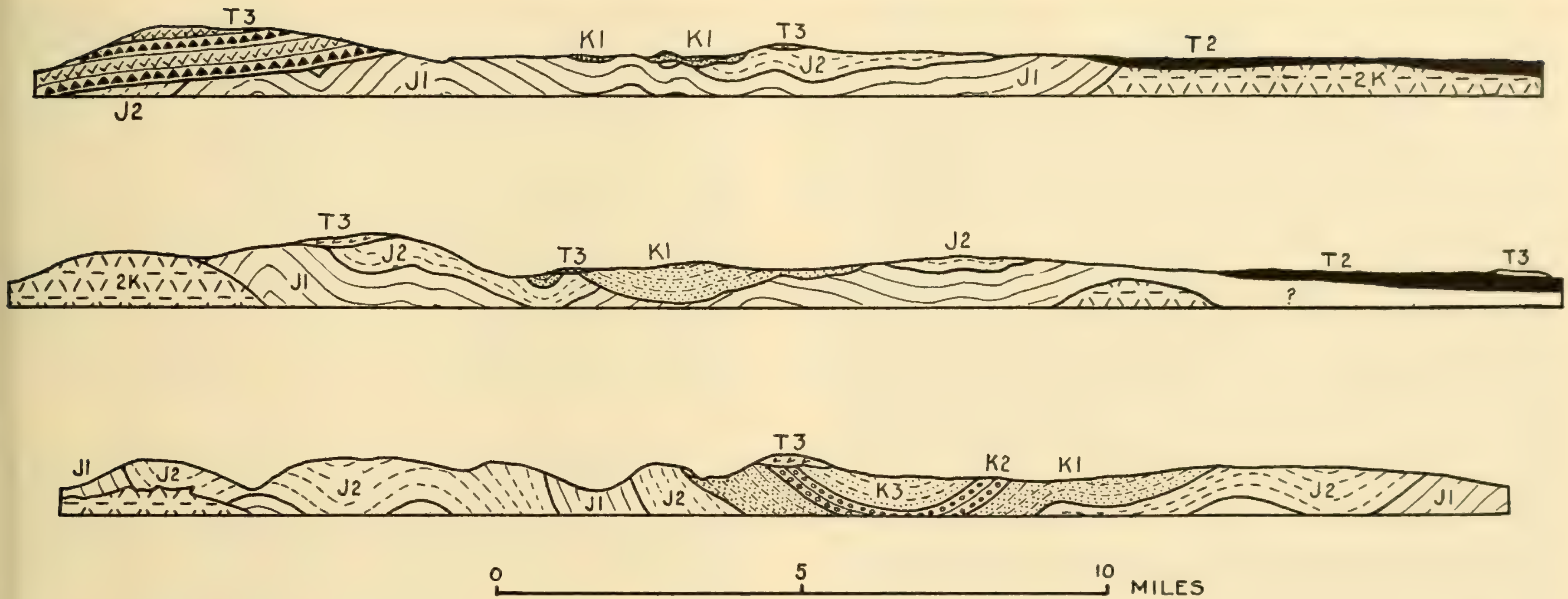


Fig. 17.23. Cross sections of the east coast of the southern part of Graham Island of the Queen Charlotte Islands, British Columbia. After MacKenzie, 1916. J1, Maude fm. (banded argillites and tuffs. Lower Jurassic or Triassic ?); J2, Yakoun fm. (basal agglomerates and minor flows. Middle Jurassic); K2, Kano quartz diorite; K1, Haida fm. (sandstone, shale, and coal. Upper Cretaceous); K2, Honna fm. (conglomerate and sanastone. Upper Cretaceous); K3, Skidegate fm. (sandstone and shale. Upper Cretaceous); T2, Skonun fm. (sandstone, shale, and conglomerate. Lower Pliocene ?); T3, Masset volcanics (basalt flows and agglomerates. Pliocene ?).

obstructed by the relatively much greater impermeability across the foliated surfaces and that portion highly metamorphosed above and adjacent to the batholith must lie below the present topographic surface, deeper down on its flank.

The gneissic structure of the batholith suggests that the magma moved upward along planes dipping steeply to the northeast and that the maximum effect of its thrust was directed against the adjoining formations on the southwest. The country rock was probably irregularly domed up to a considerable extent by the invading magma, was fractured, faulted, and stoped to some extent, and was thrust aside to a very considerable degree. There is abundant evidence in residual structures and in the composition of the resulting rocks that, locally, narrow belts of sediment were wholly incorporated in the magma through a process of reactive replacement, but this was probably not the major factor in the process of emplacement of the batholith.

Idaho Batholith. The Idaho batholith is part of the Nevadan orogenic belt, but at the same time it is closely associated with the Laramide Rockies whose building occurred at a slightly later time. Because of the complex geology around it the great pluton is treated separately in Chapter 21.

Late Cretaceous Phase

The only Upper Cretaceous deposits of the Columbia system are confined at present to a narrow belt along the northeast coasts of Vancouver and Graham Island in the Queen Charlotte group. Although the belt is narrow, the sediments have a thickness of 10,000 feet (Gunning, 1932).

The belt of Vancouver Island may be projected south-southeastward to a deposit of very thick Upper Cretaceous strata in Washington (see maps, Figs. 17.13 and 17.22). If the two were connected, as seems possible, then a narrow but deep trough formed in this area after the orogeny and batholithic intrusions of the late Jurassic and early Cretaceous phase. The trough was continuous to Graham Island and, if farther, then it must now be west of any land in southeastern Alaska and, therefore, in the continental shelf.

Near the base of the series on Vancouver Island, coarse conglomerate is found. It contains angular to subangular pebbles and boulders, on the average 2 inches in diameter but varying greatly in dimensions, of volcanic rocks, granodiorite, argillite, and quartzite. This conglomerate probably indicates the existence of a closely adjacent highland being actively elevated while the trough sank. The Upper Cretaceous strata of Vancouver Island occur in an open basin and dip about 15 degrees toward the center. The gentle folding occurred before the intrusion of dikes and stocks which are believed to be Eocene or Oligocene in age (Gunning, 1930).

Farther north on Graham Island, the Upper Cretaceous strata have been folded somewhat more intensely. See cross sections of Fig. 17.23. It is evident that a very late Cretaceous or early Eocene episode of folding affected the thick sediments of the narrow Upper Cretaceous trough. Since the history of the Eocene in nearby Washington and Oregon is chiefly one of trough subsidence, it seems best to assign the disturbance to the later Upper Cretaceous and to relate it provisionally to the Santa Lucian orogeny of the central Coast Ranges of California.

Peacock (1935) imagines that the Upper Cretaceous of Vancouver and Graham islands was once more widespread than now, and that by the close of the Cretaceous wide arms of the sea washed the margins of remnants of the once great mountain system reduced to insignificant relief. Because no Upper Cretaceous rocks are known in southeastern Alaska, it is concluded that the region there was land for the rest of the Cretaceous. Not until Eocene time did any significant subsidence occur.

It will be recalled that the Coast Ranges of California are composed mostly of the trough sediments, and the Island Ranges of southern British Columbia are only in small part latest Jurassic and Cretaceous; they are mostly the Nevadan complex. In southeastern Alaska, an offshore belt of post-batholithic Cretaceous strata may exist, however, but submerged beneath the continental shelf. See Fig. 17.21.

18.

ROCKY MOUNTAINS IN MESOZOIC TIME

TRIASSIC GEOGRAPHY

The seaways that had existed in the Paleozoic miogeosyncline were considerably changed during Mesozoic time, and a wide belt of land gradually rose in the middle of the old geosyncline to separate two troughs of sedimentation. The western trough as recounted in Chapter 17 was filled with more than 30,000 feet of interbedded sediments and volcanics and was subjected to repeated orogeny. The eastern trough was filled with marine and nonmarine beds with only a trace of volcanic material. The eastern was fairly stable with disturbance reaching orogenic proportions only in the late Mesozoic along the border of the geanticline.

The Triassic sediments were spread in the shape of a wing of a butterfly over the Rocky Mountain states (Plate 9). The site of greatest subsidence was along the east margin of the former Paleozoic geosyncline, or along the Wasatch line, where marine waters entered and considerable limestone was deposited, such as the ammonite bearing Thaynes formation of northern Utah and southeastern Idaho. The geanticline which started to emerge in Permian time became more pronounced in Triassic time, but still a wide shallow connection existed with the Pacific. It seems also that a southwestern passage to the Pacific existed. East of the marine deposits the sediments are mostly of flood-plain origin and are deep red. They are now known to extend northeasterly over part of the Williston basin. They overlapped the edges of the Ancestral Rockies but did not bury them completely.

EARLY JURASSIC GEOGRAPHY

The folio of the U.S. Geological Survey, *Paleotectonic Maps of the Jurassic System* is taken here as a guide and should be referred to for details of the distribution, thickness, and lithology of the several time divisions of the system in the United States (McKee *et al.*, 1956). Four major units are recognized which from oldest to youngest are labeled A, B, C, and D. The two oldest which include strata of Lias, Bajocian, Bathonian, and Callovian ages are represented in Plate 12. They include the well-known continental sandstone formations, Nugget, Navajo, and Kayenta, and the marine limestone and shale formations, the Twin Creek, Gypsum Springs, Lower Sundance, Sawtooth, Carmel, etc.

The Cordilleran geanticline became continuous by Early Jurassic time and joined with a large emergent area of the southwestern states and Mexico.

In Middle Jurassic time an irregular island in western Montana was uplifted. It is known as the Sweetgrass arch and stretched from Great Falls northward to the Canadian border. About and over it unconformities occur which involve the Jurassic Sawtooth, Rierdon, Swift, and Morrison formations and Early Cretaceous Kootenai formation. See Fig. 18.1. The lowest of the formations, the Sawtooth, is sandstone, siltstone, sandy lime-

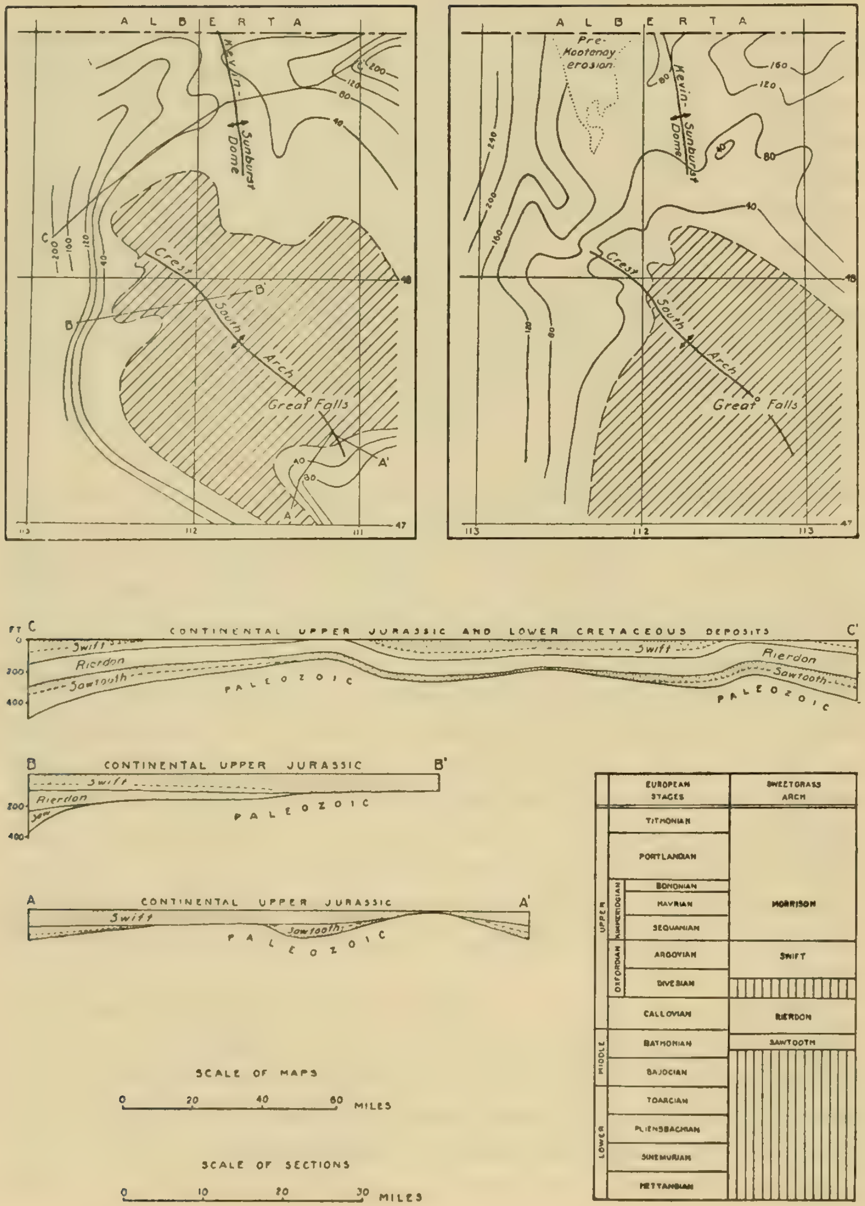


Fig. 18.1. The Sweetgrass arch in Jurassic time, after Cobban, 1945. Left, isopach map of the Sawtooth formation. Right, isopach map of the Rierdon formation. Ruled areas were exposed Paleozoic strata just before deposition of the Swift formation. The crest lines of the Kevin-Sunburst dome and the south "arch" are those of the present, and together they make up the Sweetgrass arch.

stone, sandy oolite, and shale. The medial Rierdon is largely limy shale and nodular limestone. The upper Swift is dark noncalcareous shale and flaggy glauconitic sandstone. The Morrison consists of fine-grained greenish gray clay shale and fresh-water limestone. From the structural point of view, it is significant that the Swift overlies the other formations unconformably (see cross section, Fig. 18.1), and indicates that the Sweetgrass arch rose in early Late Jurassic time in about the same position and with the same detail as that produced by later Laramide movement. The arch consisted of two domes, a northern and a southern; the southern one was the site of greatest uplift and erosion. It is also believed that the southern dome rose gently and remained above sea level during the deposition of the Sawtooth and Rierdon formations, just preceding early Upper Jurassic uplift.

The Morrison was deposited conformably on the Swift; following Morrison time, the arch was again elevated slightly, and several low anticlines and synclines were formed. Subsequent erosion removed part of the Jurassic beds from the crests of the anticlines. Erosion was most pronounced on an anticline extending approximately north-south through the north dome (Kevin-Sunburst). The Morrison, the Swift, and part of the Rierdon were removed along the crest of the anticline and westward for an unknown distance. Over this area the earliest Kootenai sands and gravels (Cut Bank sandstone) were deposited, while the area east and south continued to undergo erosion. It was not until Sunburst time (a sandstone member of the Kootenai) that the entire area received sedimentation.

By the close of Mid-Jurassic time (Callovian) the Sweetgrass arch had spread as a land area of very low relief to include central and western Montana and northwestern Wyoming. See Plate 12.

The Ancestral Rockies had been overlapped still more in Early and Mid-Jurassic time, and during the Late Jurassic were entirely buried save for a few peaks in the Front Range of Colorado.

EARLY AND MID-CRETACEOUS OROGENY

The Fernie strata in Alberta appear to grade into the overlying Kootenay formation of Lower Cretaceous age. The Kootenay consists of alternating sandstone and dark shale with many coal beds, perhaps all of nonmarine origin, and decreases in thickness from west to east. Its greatest thickness is 5000 feet. The presence of thick sandstone and conglomerate beds in the Kootenay is indicative of further uplift of the land to the west, and the presence of granite pebbles in the conglomerates indicates that erosion and differential movement by Kootenay time had so far proceeded as to lead to the uncovering of deep-seated plutonic masses.

Near the south end of the Wasatch Mountains in the Cedar Hills, Gunnison plateau, and Sanpete Valley are immense, coarse deposits probably of early Late Cretaceous age. They make up the basal part of the Indianola group (Spieker, 1946). See Fig. 22.16 and Chapter 22. Since the conglomerates, sandstones, and shales, together with some higher fossiliferous marine beds, are a lithologic unit, Spieker believes that all the deposit is a consecutive response to an uplift, and therefore that the orogeny occurred at the beginning of late Cretaceous time. Because of the thick deposit of Mid-Cretaceous age in the Cedar Hills, and the information obtained there about the disturbance, the orogeny will be called after them, namely, the Cedar Hills orogeny.

The clastics of the Indianola group are coarsest toward the west in the Cedar Hills (Schoff, 1937) and the Gunnison plateau. They grade eastward into the Mancos shale at the east front of the Wasatch plateau. The greatest thickness known is 15,000 feet in the Cedar Hills. The belt of intense deformation lay west of the Cedar Hills, because in the Cedar Hills the conglomerates rest conformably upon the underlying Upper Jurassic shales (Spieker, 1946).

The belt of the Cedar Hills orogeny must have extended from southern Nevada northward through Utah to eastern Idaho. In southern Nevada, the Overton fanglomerate in the Muddy Mountains is of early late Cretaceous age (Hewett, 1931), and rests in angular unconformity on folded and thrust-faulted Mesozoic rocks, the youngest of which are Jurassic in age (Longwell, 1928, 1936). The Overton fanglomerate and the angular unconformity are believed to mark the Cedar Hills orogeny in southern Nevada, and the inference has been made that the belt of orogeny extended continuously between southern Nevada and central Utah.

North of the Cedar Hills in north-central Utah, a coarse conglomerate, the Kelvin, is probably the equivalent of the lower Indianola conglomerates. It is about 200 feet thick and grades eastward into finer sediments. The uplift lay immediately west of the present Wasatch Mountains, and Permian cherts and Pennsylvanian quartzites in the uplift furnished most of the pebbles of the conglomerate. The site of conglomerate accumulation became a trough of subsidence, and in the Colorado epoch of late Cretaceous time, over 5000 feet of strata collected in it. Volcanoes nearby emitted dust which collected as tuff in the lower part of the sequence (the Aspen formation); then sandstones with numerous oysters and fresh-water shales and sandstones with coal seams accumulated alternately. Several conglomerates in the Colorado series mark continued unrest to the west. See the paleotectonic map, Plate 12.

Mansfield (1927) believes that the Lower Cretaceous Gannett group in southeastern Idaho, with its several coarse sandstones and conglomerates, signifies a sharp uplift in the land to the west as a reflection of the intense Nevadan orogeny still farther west. Probably this uplift in the Utah trough area was a forerunner to the main orogeny which resulted in the deposition of about 3000 feet of coarse debris of early Late Cretaceous age, the Wayan formation (Read and Brown, 1937) unconformably on the Gannett.

About 200 miles northwest of southeastern Idaho in southwestern Montana, a Cretaceous sequence is present, but has not yet been well worked out. In places below beds of Aspen (Colorado epoch) aspect, and above the Lower Cretaceous Kootenai clastics is a pebble and cobble conglomerate. Although these beds may be part of the Kootenai, fossil evidence is lacking and they may be early Late Cretaceous. If so, the belt of Mid-Cretaceous orogeny may have extended northward to western Montana.

Since no Jurassic and Cretaceous beds were deposited in the Mesozoic geanticlinal area along the east side of which the Cedar Hills orogeny oc-

curred, the folds and faults there are all in Paleozoic strata, and therefore the date of the folding cannot be fixed except as post-Paleozoic. The main orogenic events in the eastern trough came in Late Cretaceous time and during the Paleocene and Eocene, and therefore the folds and faults in the Paleozoic strata of the geanticline immediately to the west have generally been accredited to these later orogenies. However, the work of Nolan (1935) in the Gold Hill mining district of western Utah is especially significant in making clear the complexity of deformation in the geanticlinal area. There the structural history is characterized by at least four and possibly five phases of folding and faulting, each phase composed of an initial stage in which compressive forces were active and a final stage in which normal faulting was dominant. The first two phases predate the Eocene by a long interval of erosion and are regarded provisionally as Cretaceous by Nolan. It is probable that they are related to the Nevadan and post-Nevadan Cretaceous disturbances to the west (see chart, Fig. 17.2 and 17.7) and to the sinking of the Utah trough during the time that the Indianola, Kelvin, Aspen, and Frontier and other formations were deposited in it. The map, Plate 10, shows the crust intensely affected in the Sierra Nevada region in late Upper Jurassic time, while the area on the east was only epeirogenically uplifted. During Cretaceous time, the reverse seems to have been true. The Sierra Nevada region was one of gentle emergence, and the eastern part was probably orogenically deformed.

In conclusion, there is no evidence to preclude the generalization that the most intense disturbance in the landmass just west of the trough was localized opposite the area of greatest subsidence, which also coincides with the central part of the arcuate pattern. See especially Plates 11 and 12.

19.

LATE CRETACEOUS AND EARLY TERTIARY ROCKY MOUNTAIN SYSTEMS—THE LARAMIDE OROGENY

DEFINITION OF LARAMIDE OROGENY

Geologists to date in the Rocky Mountains have discovered a succession of dynamic events through late Mesozoic and Tertiary time. At first, a single, rather violent orogeny was visualized, but now numerous unconformities, coarse conglomerates, and structural relations attest a condition of unrest in the general Rocky Mountain region from middle Mesozoic to the present. The single and intense orogeny in the Rocky Mountains which they visualized was called the Laramide Revolution, and this was supposed to have occurred precisely at the close of Cretaceous time, or at the beginning of Tertiary time.

Some geologists advocate dropping the term Laramide because of the many recognized deformational pulses and the current concept that crustal deformation is continuous. Limits cannot logically be set, they contend. The writer believes, however, that since the usage is so deeply ingrained in the literature that it is better to try to define the term arbitrarily, and furthermore, finds the attempt helpful and not confusing. For the purposes of this book the following nomenclature will be used:

Orogenic events during Eocene time—Late Laramide
Orogenic events during Paleocene time—Mid-Laramide
Orogenic events during Montana time—Early Laramide

Any orogenic phases older than Montana or younger than Eocene will not be called Laramide, and, where desirable, new orogenies will be defined. The Cedar Hills orogeny of central Utah of Colorado age falls in this category. The disturbances in late Mesozoic time were generally precursory to the climatic ones of the very Late Cretaceous or the Early Tertiary.

BELTS OF DEFORMATION

Major Divisions

The map, Fig. 19.1, has been prepared to show the mountain systems of the Laramide orogenic belts. Two major divisions of the systems have been pointed out in the literature, namely, a western composed of ranges formed of the thick sediments of the Paleozoic and Mesozoic troughs, and an eastern composed of ranges and intermontane valleys formed of the shelf sediments and the crystalline basement complex (Fig. 19.2). The generalization needs scrutiny from both a spatial and time aspect, else a number of misconceptions will arise. This will be done in the following several chapters.

General Characteristics

Thrust faults and folds are the most characteristic structures of the Laramide Rocky Mountains. In the eastern division great asymmetrical anticlinical ranges dominate. Those that have been uplifted so much

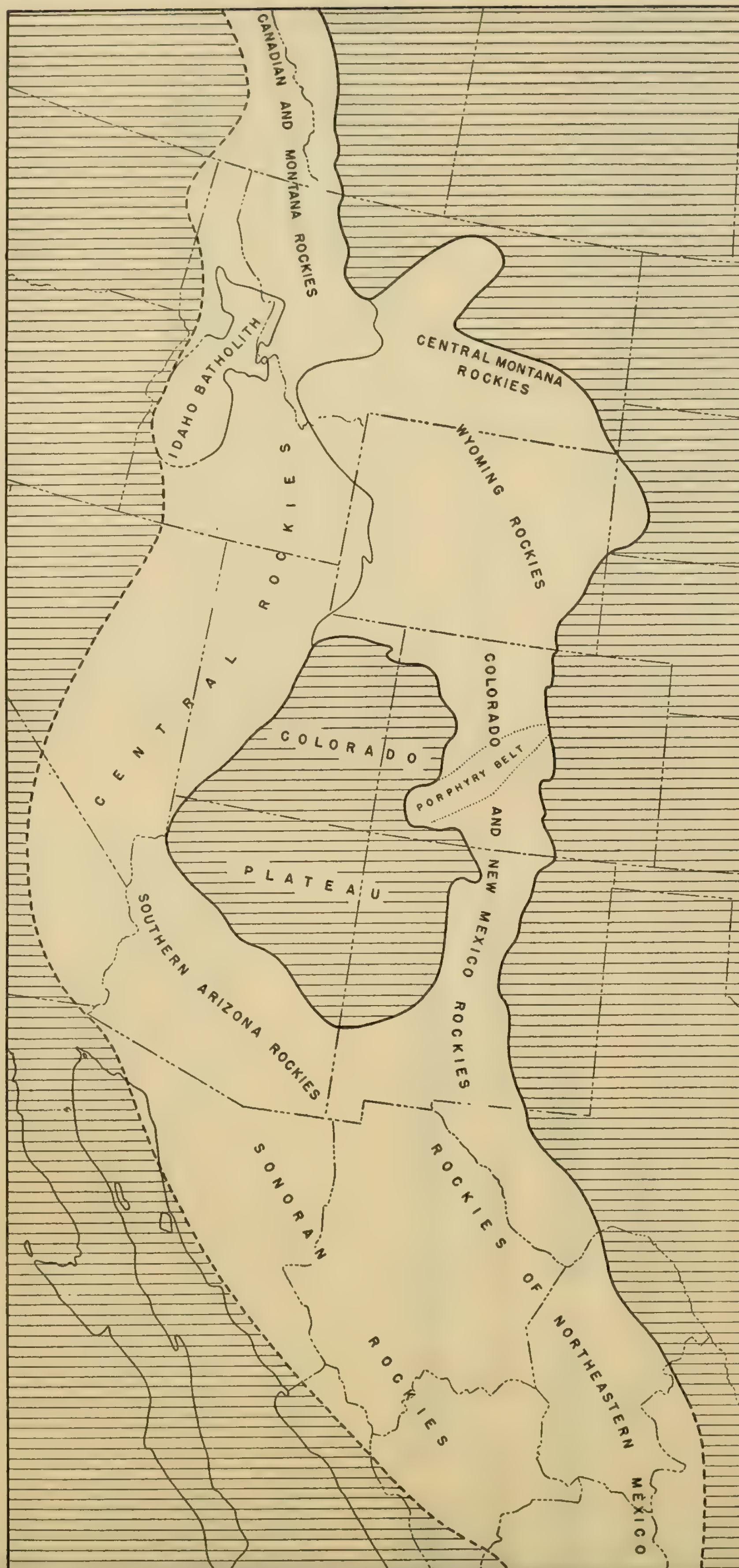


Fig. 19.1. Divisions of the Laramide Rockies which are treated under separate chapters.

that Precambrian cores show extensively are marked by thrusts on the steep flank (Fig. 19.2). Such structures are here interpreted as primary upthrusts and gravity slide phenomena.

Many of the thrust faults of the western division are low angle and define sheets that have moved horizontally considerable distances. Some of the thrust sheets are folded by later movements. The thrust sheets were rather thin and escaped regional metamorphism. In fact, the rocks involved in the Laramide orogeny are characterized by an absence of metamorphism, except perhaps some of the deeper Proterozoic strata. This distinguishes them from the rocks of the Nevadan orogeny. It will be recalled that the Nevadan in the batholithic belt is characterized by isoclinal, nearly vertical folds, as well as flow cleavage. Isoclinal folding is rare in the Laramide Rockies. The Nevadan is characterized by great batholiths. Aside from the Idaho batholith, the Laramide orogenic belt has few plutons large enough to be called batholiths; its intrusions are mostly stocks, but the stocks exist in considerable number. The Nevadan developed in sediments of the eugeosyncline, the Laramide in sediments of the miogeosyncline and shelf.

Canadian and Montana Rockies

The Canadian and Montana Rockies consist of a mainland assemblage of geosynclinal sediments of late Proterozoic, Paleozoic, and Mesozoic age, cast into a great imbricate series of thrust sheets. The Proterozoic rocks of western Montana form an extraordinarily thick group of clastic sediments, known as the Belt series. Originally clays, sands, and marls, they have been metamorphosed to argillites, quartzites, and impure sideritic marbles and limestones. They are at least 50,000 feet thick near Missoula. The Paleozoic rocks are dominantly limestones, and are nearly 7000 feet thick. The Madison limestone of Mississippian age is about 2000 feet thick, and forms steep cliffs and canyon walls in many of the ranges southeast of Missoula. The Mesozoic rocks are about 7000 feet thick and are dominantly shales, with some limestone, sandstone, and conglomerate. Consult tectonic and geologic maps of the Permian, Triassic, Jurassic, and Cretaceous, Plates 8 to 12 for information on the deposition and distribution of the various stratigraphic systems, and Chapters 5 and 6 for isopachs.

The Beltian has not been reported to be as thick in Idaho and Utah as in western Montana, and north of the border it also seems to be thinner. There, it crops out almost entirely west of the Rocky Mountain trench and leaves the main Canadian Rockies to be composed mostly of Paleozoic strata. The Cambrian thickens to over 15,000 feet along the Alberta–British Columbia boundary, and most of the scenic ranges there are sculptured in it.

Very few intrusions occur east of the Rocky Mountain trench north of the Idaho and Boulder batholiths. Large sheets of diorite and gabbro split the Beltian rocks in places, and one near the Canadian border is identified as a lava flow and is called the Purcell lava. The sills and flows have been deformed with the Beltian strata.

The igneous intrusions are very abundant and voluminous in west-central Montana, and are composed chiefly of quartz monzonite and diorite.

Central Rockies

The central Rockies consist of folded and thrust-faulted Paleozoic strata in their western part and Proterozoic, Paleozoic, and Mesozoic along their eastern margin. The Mesozoic sediments were especially thick in places, and a number of episodes of compression occurred from mid-Cretaceous to early Oligocene. The structures of some of the episodes trend discordantly to those of others. Thick, coarse conglomerates mark the orogenies and, being deformed themselves, add to the complexity.

A review of the tectonic maps of the Permian, Triassic, Jurassic, and Cretaceous, Plates 8 to 12, will impress one with the fact that the Laramide trough zone of orogeny, especially in Utah, embraced parts of two major elements, the Cordilleran intermontane geanticline and the Mesozoic trough. The sinking of the Permian trough in Utah started a series of subsidences that followed generally one on top of the other until the Laramide orogeny. The total accumulation of sediments of the Permian and Mesozoic, therefore, has been isopached, and the basin is charted on a map so as to compare with the Laramide deformational belt. The map, Fig. 19.3, shows the extent to which the Laramide belt cut into the

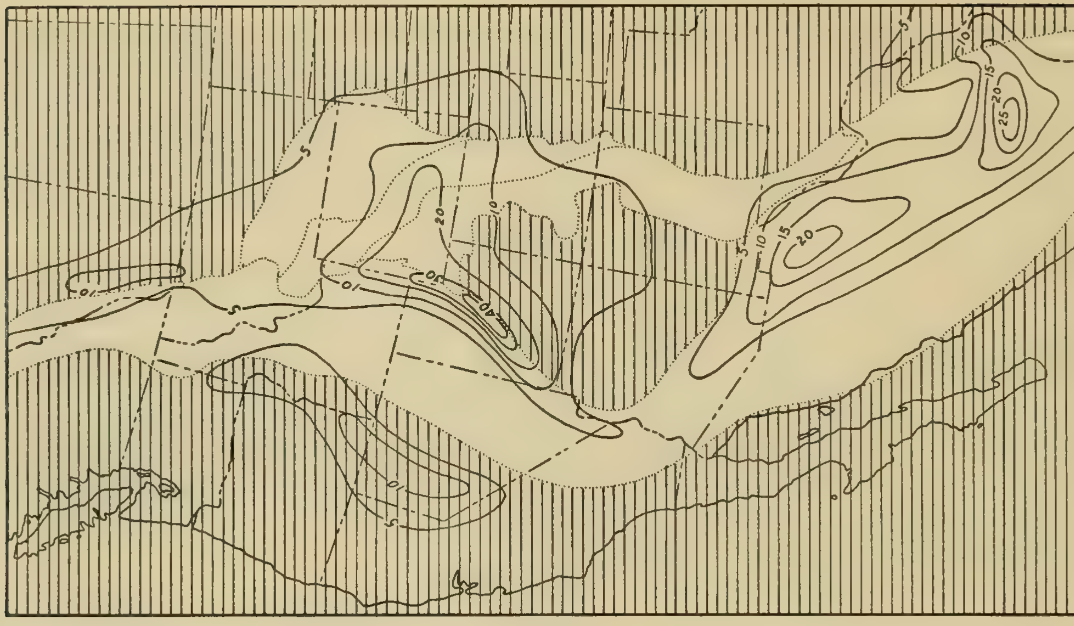
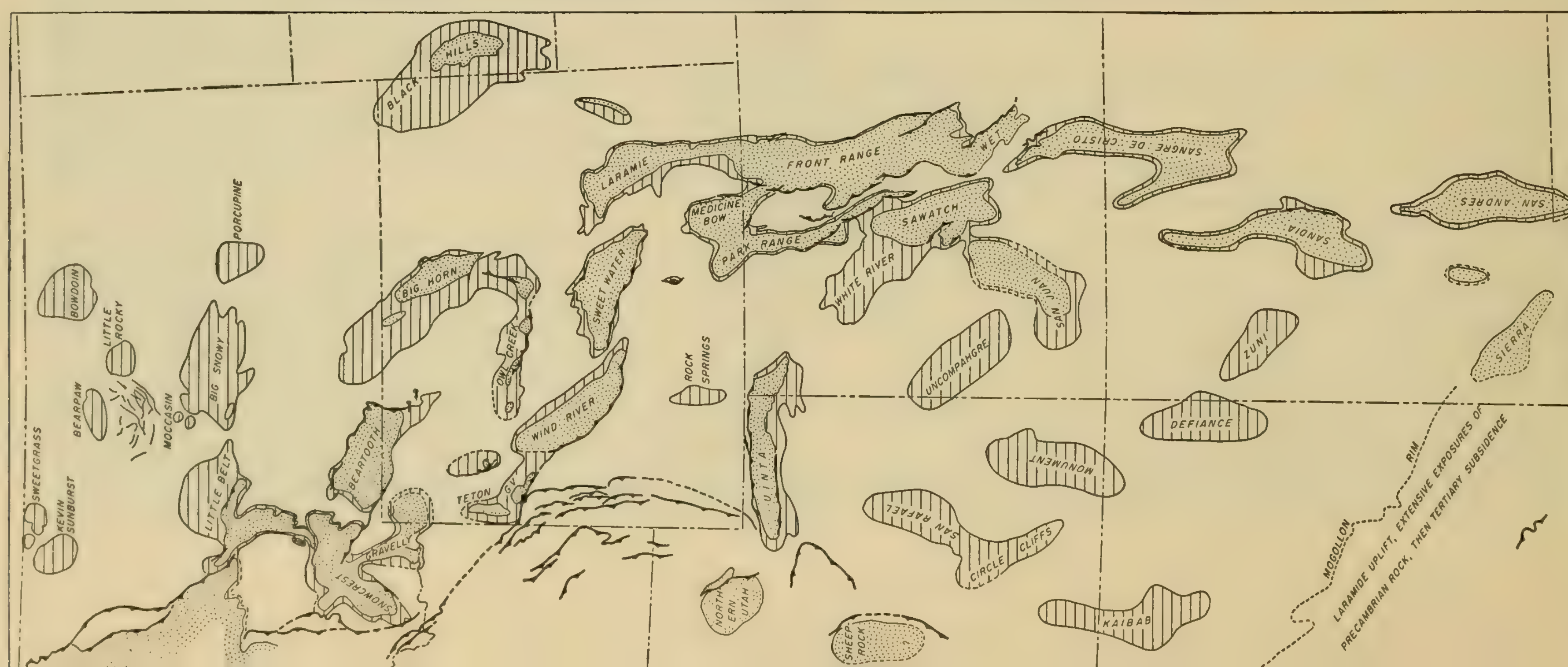


Fig. 19.3. Permian and Mesozoic basins (contoured) and the belts of the Laramide orogeny (white). The isopach of the combined Permian, Triassic, Jurassic, and Cretaceous basins are numbered in thousands of feet. The Permian basin in western Oregon is also shown.



ig. 19.2. Laramide uplifts of the shelf province of the Rockies. Stippled areas are

Permian and Mesozoic trough sediments and the extent to which it overlapped the Paleozoic sediments of the geanticlinal area.

Precambrian rocks are not exposed at the surface in many ranges, but from central Utah northward to western Montana, those of Proterozoic (Beltian) age become increasingly widespread. In western Montana, most of the Laramide Rockies are in the Beltian strata, and this zone extends to the northwest in eastern British Columbia. The crystalline complex, supposedly everywhere older than the Beltian, is exposed in the trough zone only in the Wasatch and Raft River Mountains. In the shelf ranges the opposite is true; the crystalline complex is exposed in many of the cores of the ranges. The extent of the Beltian trough, as well as can be determined, is shown in comparison with the Laramide belts of deformation in Fig. 19.3 and will be referred to later.

Thrust faults dominate the structure in the trough zone. The overriding sheets of greatest displacement and shallowest dip moved mostly eastward, but several thrusts, especially in Montana and Canada, have moved westward.

Southern Arizona Rockies

The southern Arizona Rockies consist largely of Precambrian rocks of several ages, both igneous and sedimentary. The ancient rocks were veneered with a thin Paleozoic cover, and in places with thin Triassic and Jurassic strata. The Cretaceous Mexican geosyncline extended northwestward into the southernmost part of Arizona, and its strata are there thrown into folds and thrust sheets. Part of the Cretaceous accumulations were lavas. See the Paleotectonic maps for the details of the setting for the Laramide orogeny. Intrusive rocks of Laramide age are abundant, and they are associated with valuable ore deposits. A succession of volcanic episodes spread through the Tertiary, and probably some are early enough to be considered Laramide.

Wyoming Rockies

The Wyoming Rockies consist in part of the shelf facies of Paleozoic, Triassic, Jurassic, and Lower Cretaceous rocks, in part of thick clastic deposits of Late Cretaceous age and in part of Beltian (?) and pre-Beltian crystalline rocks. The most conspicuous ranges are sculptured out of great asymmetrical anticlines in which the Precambrian rocks are exposed in the cores. The Black Hills, Big Horn, Laramie, and Wind River ranges are eroded in such folds. The anticlines are asymmetrical to the extent of overturning and thrusting in places, and began to rise in Late Cretaceous time, while the broad basins between sank and received thousands of feet of sediments. Examine the paleotectonic map of the Late Cretaceous. The thick Upper Cretaceous sediments are generally involved in late phases of the compressional orogeny. Paleocene and Eocene sediments have accumulated in the basins to considerable thicknesses in places, and certain phases of deformation marginal to the basins have deformed them also.

The northwest corner of the state of Wyoming became the site of considerable volcanic activity in middle and late Eocene time, and the pyroclastics and lavas of the Absaroka Range and Yellowstone Park were mostly exuded at that time.

Central Montana Rockies

The central Montana Rockies consist of a general east-west assemblage of monoclinal flexures, domes, and belts of *en echelon* faults. Numerous bodies of igneous rocks consisting of stocks, laccoliths, radiate dike systems, and various extrusions lie in a belt approximately transverse to the sedimentary rock structures.

During the Beltian epoch of the Proterozoic, a trough extended eastward into central Montana and may have predetermined the location of the central Montana Laramide structures. See map, Fig. 19.4. Also, in Mississippian times a broad east-west basin through central Montana subsided and received over 2000 feet of beds, appreciably more than on either side. See paleotectonic map, Plate 5. This basin, like the Beltian, may have helped to determine the position that the later Laramide structures took, or else the coincidence in space of all three means that some deep-seated influence has been at work repeatedly from Beltian times to Laramide.

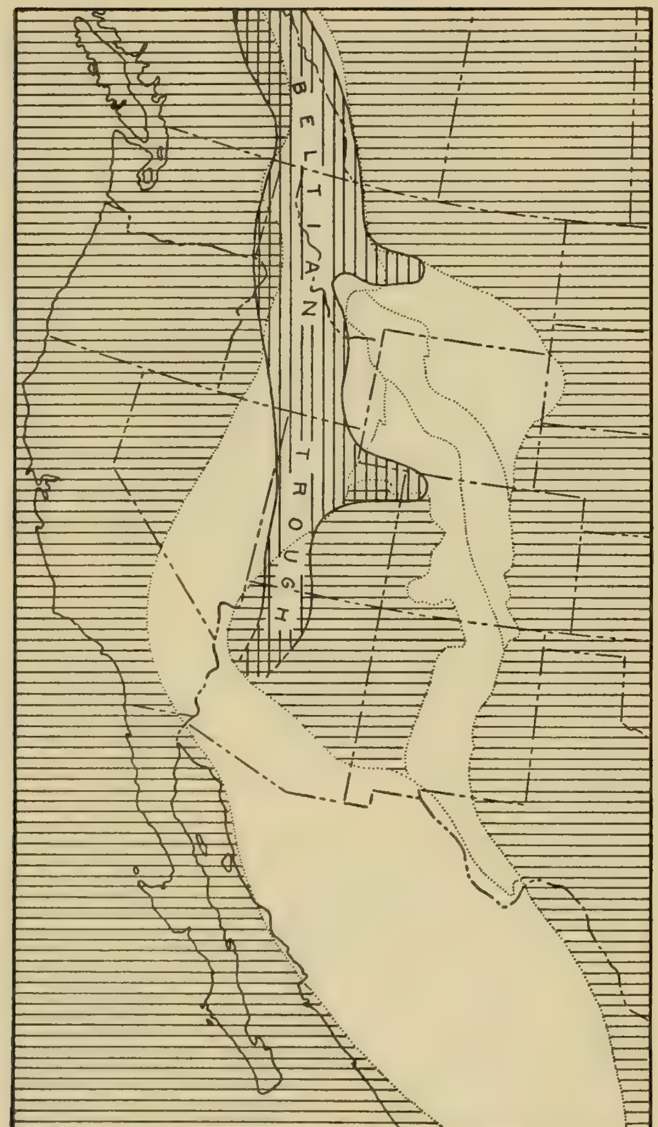


Fig. 19.4. Beltian trough (vertically ruled) and the belts of the Laramide orogeny (white).

Colorado and New Mexico Rockies

The Ancestral Rockies of Pennsylvanian and Permian age were gradually buried by Triassic, Jurassic, and Cretaceous sediments, and it was upon this crustal make-up that the Laramide belt of deformation was superposed in Colorado and New Mexico. The back of the ancestral Colorado Range was broken, and two modern ranges were created, both with subparallel elements such that the older range seemed to have exerted some control over the younger. The western half of the ancestral range with a thin sedimentary veneer on the Precambrian crystallines developed a number of thrust sheets. A transverse prophyry belt carries numerous stocks and much ore.

The Laramide belt of deformation in New Mexico was a narrow one through the central part of the state, and aside from the common north-south orientation of both the Ancestral and Laramide Rockies, their relation seems to be a matter of chance, viz., the younger ranges rose in part where the older ones stood and in part in the sites of the older basins.

The Laramide structures are large asymmetrical anticlines like those in Wyoming, with gravity slide thrusting on the steep flanks. Graben or rift faulting broke through the Laramide uplifts in a north-south zone in Late Cenozoic time. Much Tertiary volcanism occurred in Colorado and New Mexico.

Rockies of Northeastern Mexico

The Laramide system of northeastern Mexico includes the El Paso–Rio Grande thrust belt, the Sierra Madre Oriental, the Sabinas foothill belt, and the Parras synclinorium. The Pennsylvanian Marathon and Coalhuila systems, the Permian ranges, platforms, basins, and shelves of the Marathon foreland, and the late Mesozoic Mexican geosyncline are the foundational elements upon which the Laramide structures were superposed. The strata of the Mexican geosyncline were generally closely folded lengthwise of the basin, the thin veneer on the old Coalhuila system—the Coalhuila peninsula—was domed broadly and locally flexed, and the basin beds along the east side of the peninsula were folded and thrust eastward. The Parras trough at the south end of the peninsula was intensely compressed from south to north, and tight east-west folds and

some thrusts were formed. The Pennsylvanian and Permian Marathon and Coalhuila systems are, therefore, thought to have extended considerable control over the later Laramide structures.

Sonoran Rockies

The Sonoran Rockies include three geomorphic provinces, namely from east to west, the Sierra Madre Occidental, the parallel ranges and valleys, and the Sonoran desert. From coarse conglomerates along the western margin of the Mexican geosyncline, it is clear that the orogenic belt in western Sonora continued active, and at least twice in late Cretaceous time rose sharply and crowded the sediments eastward. The early Laramide structures thus created are obscure, first for lack of field work, and second because other younger orogenies have been superposed, and much Tertiary lava covers them.

The Paleozoic and Mesozoic sediments are of a mainland assemblage, as far as known, and the volcanic assemblage of the Pacific border systems is absent. On the other hand, the Permian beds of the Coahuila peninsula, a considerable distance to the east, have much volcanic material.

Colorado Plateau

The Colorado Plateau is a rudely circular and lesser deformed part of the crust within the broad zone of Laramide orogeny. A sedimentary veneer of about 6000 to 10,000 feet overlies a Precambrian basement, and over much of the Plateau the strata are nearly flat. Several large monoclinal flexures of Laramide age break the monotony of the flat-lying beds.

The monoclines are the steep flanks of asymmetrical anticlines in size much like those of Wyoming, Colorado, and New Mexico, but with about half as much vertical uplift. Consequently, it is believed, no thrusts have developed through gravity gliding as in the ranges where the Precambrian cores are so broadly exposed, and topographic relief is so much greater.

In post-Laramide time, the Colorado Plateau became the site of considerable intrusive and extrusive igneous activity, but it must not be inferred that the igneous activity was confined to the Plateau. It was equally pronounced in the more severely deformed belts to the east and west. The Plateau includes part of the Ancestral Rockies in its eastern part and contains along its western edge some Beltian (?) strata, but most of the Precambrian is a pre-Beltian crystalline complex.

RELATION OF BELTS OF DEFORMATION TO CRUSTAL CONSTITUTION

The outer ranges of the Rockies were all developed in the shelf zone of the westward-lying Paleozoic Cordilleran geosyncline, and shelf conditions of deposition continued through the Triassic and Jurassic. But with the coming of Cretaceous time, rather thick masses of sediments accumulated locally over the former shelf, particularly in Upper Cretaceous basins incident to the early uplift of the Wyoming, Colorado, and New Mexico Laramide ranges.

The belt of deformation in the shelf extends into the region of the Devonian Transcontinental Arch without effect. The thick Proterozoic metasediments, perhaps all of Beltian age, are shown as well as possible in relation to the Laramide belts of deformation in Fig. 19.4, and no striking coincidence is noted, except locally, perchance in the Uinta Range and the mountains of central Montana. If the relations as depicted are correct, then only one conclusion seems warranted, namely, that the belts of deformation are due to deep-seated causes, not influenced particularly by deeply filled troughs or basins, nor by the crystalline basement with a thin veneer of sediments.

20.

CANADIAN AND MONTANA ROCKIES

MAJOR SYSTEMS OF CANADIAN CORDILLERA

The Geological Survey of Canada classifies their great Cordillera into the western and eastern regions, and the western region is further subdivided into the western system of coastal ranges and the interior system of plateaus and ranges. The eastern Cordilleran region is spoken of as the eastern system. Examine map, Fig. 20.1.

Western System

The western system, which here will include the coast ranges and islands of southeastern Alaska, is basically Nevadan in its geological complexity, and its Mesozoic history has already been described. Its Tertiary

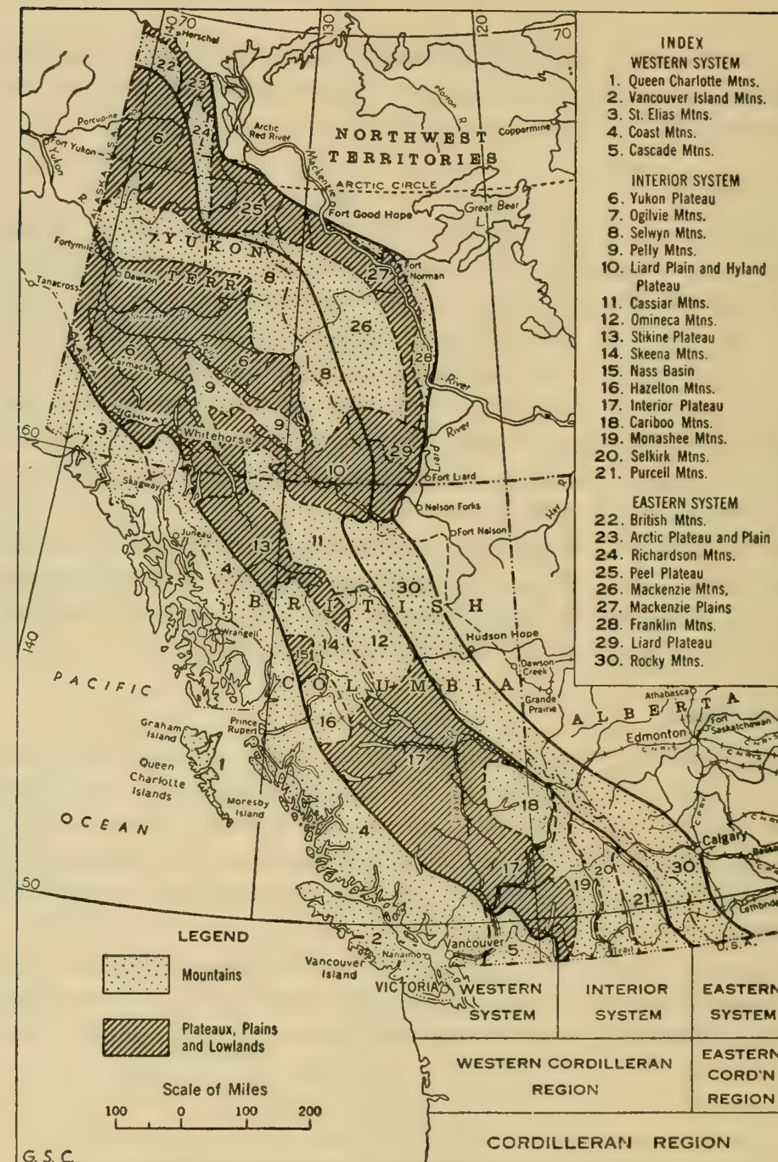


Fig. 20.1. Physiographic divisions of the Canadian Cordillera. Reproduced from Bostock *et al.*, 1957.

history will be considered in a later chapter. Quoting from Lord *et al.* (1947):

The western system includes the St. Elias, Coast, Cascade, and Vancouver Island Mountains. The St. Elias Mountains occupy an area in the extreme northwest corner of British Columbia and adjacent southwestern Yukon. They are the highest in Canada, extremely rugged, and in large part covered by an icefield. The elevation of Mount Logan, the highest peak in Canada, is 19,850 feet, and other peaks exceed 15,000 feet.

The Coast Mountains occupy a belt 100 miles wide and 1,000 miles long, and border the Pacific coast from Yukon southeast almost to the International Boundary at the 49th parallel. They rise abruptly from the sea, and towards the axis of the range are characterized by an almost unbroken succession of bare, rugged peaks and saw-toothed ridges rising to elevations from 7,000 to more than 13,000 feet. Alpine glaciers and icefields are common, and in a few places in the northern half of the range valley glaciers extend to sea-level. The range is crossed by a number of deep river valleys, and its western margin is penetrated by numerous, long, narrow fiords continued inland by deep U-shaped valleys.

The Cascade Mountains project into Canada from the State of Washington and are more than 100 miles wide where they cross the border. They lie on the east side of lower Fraser River Valley, which separates them from the Coast Mountains, and extends as far north as Thompson River. Many of the higher peaks and ridges near the International Boundary attain elevations between 7,000 and 8,500 feet; and they are fully as rugged as those of the adjacent Coast Mountains, and, like them, hold many alpine glaciers.

Mountains occupy most of Vancouver Island and culminate, in the central part, in peaks 5,000 to 7,000 feet or more above sea-level. The western side of the island, like the western side of the Coast Mountains, is characterized by an intricate set of fiords and by heavily timbered rocky slopes that rise abruptly from the sea to heights of several thousand feet. A lowland as much as 10 miles wide borders the east coast.

Central System

The central system, like the western, is for the most part fundamentally Nevadan in its geology. Its present geomorphic characteristics are adequately described by Lord *et al.* (1947). Referring again to the map of Fig. 20.1 and quoting from them:

The central system, composed of dissected plateaux and scattered mountain ranges, occupies a belt that averages more than 200 miles wide and extends southeast from the Alaska Boundary at Yukon River to the southern boundary of British Columbia at Okanagan River. In Yukon it includes the Yukon Plateau and Ogilvie, Selwyn, Pelly, and other mountains. In British Columbia north of latitude 54 and 55 degrees it includes Cassiar and Omineca Mountains, Babine and Bulkley Mountains, and Stikine Plateau. In the southern part of the province, it comprises the Interior Plateau and Cariboo, Monashee, Selkirk, and Purcell Mountains.

Yukon Plateau in Canada includes much of the drainage basin of Yukon River and, commencing in northern British Columbia near Atlin and Teslin Lakes, extends northwestward through Yukon and thence westward into Alaska. It has been deeply dissected by a drainage system whose main channels are several thousand feet deep, and the once gently rolling upland has been broken into a series of high, flat-topped hills and ridges. Ogilvie and Selwyn Mountains border it on the north and northeast respectively, and to the southeast the plateau ends against Pelly Mountains.

Little is known about Ogilvie and Selwyn Mountains. The former, with bordering peaks as high as 7,000 feet, extend easterly from the Alaska boundary, near latitude 65 degrees, for 150 miles. There they join Selwyn Mountains, which form the northeast rim of the Yukon Plateau and stretch nearly 400 miles southeasterly to end in low country east of Frances River near latitude 61 degrees. Selwyn Mountains rise from the Plateau along an irregular front, and are broken into groups of mountains by broad valleys and other depressions. Probably a few peaks are more than 10,000 feet above sea-level, and many rise to elevations in excess of 7,000 feet. Selwyn Mountains are bordered on the northeast by the Mackenzie Mountains of the eastern physiographic subprovince.

Pelly Mountains form a triangular area in the southern part of the Yukon Plateau, with corners near Teslin Lake, Frances Lake, and Pelly River at longitude 135 degrees. They include Glenlyon, Pelly, and Big Salmon Ranges, and rise from adjacent plateau areas through border areas characterized by long, smooth-topped spurs and dissected tablelands. The highest peaks of the main unit, the rugged Pelly Range, may be more than 8,000 feet above sea-level, and hold a few small alpine glaciers.

Cassiar and Omineca Mountains constitute a continuous belt stretching 450 miles northwesterly from near Takla Lake into Yukon, and extending 50 to 75 miles west from Finlay and Parsnip Rivers. These mountains comprise a great number of ranges separated by broad, transverse and longitudinal valleys several thousand feet deep. The higher peaks and ridges range in elevation from 6,000 feet to more than 8,000 feet. Permanent ice is confined to rather small, scattered, alpine glaciers.

Babine and Bulkley Mountains and their northerly extensions occupy an area of more than 20,000 square miles, bounded on the east by Cassiar and Omineca Mountains, on the south by the Interior Plateau, on the west by the Coast Mountains, and on the north by Stikine Plateau. Bulkley and Babine Mountains lie on either side of the northwesterly trending Bulkley–upper Skeena Valley. They comprise many individual mountains or mountain groups isolated by wide low areas or great valleys. Most peaks are highly dissected, and some rise more than 7,500 feet above the valleys.

Stikine Plateau occupies much of the drainage basin of Stikine River east of

the Coast Mountains: on the north it joins Yukon Plateau between Atlin and Teslin Lakes, and elsewhere is bounded by the northerly extensions of Babine and Bulkley Mountains or by Omineca and Cassiar Mountains. Its gently undulating surface averages 4,000 feet or more above sea-level, and is dissected into a number of smaller plateaux by the larger stream and river valleys.

The Interior Plateau stretches from Bulkley, Babine, and Omineca Mountains approximately 500 miles southeasterly to the International Boundary. At its north end it extends from the Coast Mountains 200 miles east to the Rocky Mountains. Toward the south it becomes progressively restricted by the Cascade Mountains, on the west, and by Cariboo and Monashee Mountains, on the east, and at the Boundary near Okanagan and Kettle Rivers is less than 50 miles wide. This great plateau region, with a general elevation of 3,000 to 4,000 feet is composed of a succession of plateau surfaces interrupted by the deeply cut valleys of a drainage system whose main channels lie 1,000 feet or more below the remnants of the upland surface.

Cariboo, Monashee, and Purcell Mountains form a mountain group within a triangular area between the Interior Plateau on the west and the Rocky Mountain Trench on the east; the apex is in the big bend of Fraser River, and the base at the International Boundary. The various members of the group are separated by deep valleys or trenches trending northward and northwestward. Selkirk Mountains are exceedingly rugged, with summits rising to elevations of 11,000 feet and more above sea-level.

Eastern System

The eastern system, or the Canadian and Montana Rockies, is the subject of the present chapter because it is basically Laramide in origin. It

. . . includes Richardson, Mackenzie, Franklin, and Rocky Mountains, and intervening plateau and plain areas.

In British Columbia the eastern and central systems are separated by the Rocky Mountain Trench, a great trough that extends northwesterly from the International Boundary nearly to the southern boundary of Yukon, and includes aligned parts of Kootenay, Columbia, Fraser, Parsnip, and Finlay Rivers. The boundary between these systems is less well defined beyond the northern end of the trench; it enters Yukon near longitude 126 degrees, extends northerly into Northwest Territories, and swings to the northwest between Selwyn Mountains on the southwest and Mackenzie Mountains on the northeast to re-enter Yukon near latitude 65 degrees, and thence proceeds northwesterly on a sinuous course to pass west of Richardson Mountains and enter Alaska near latitude 69 degrees.

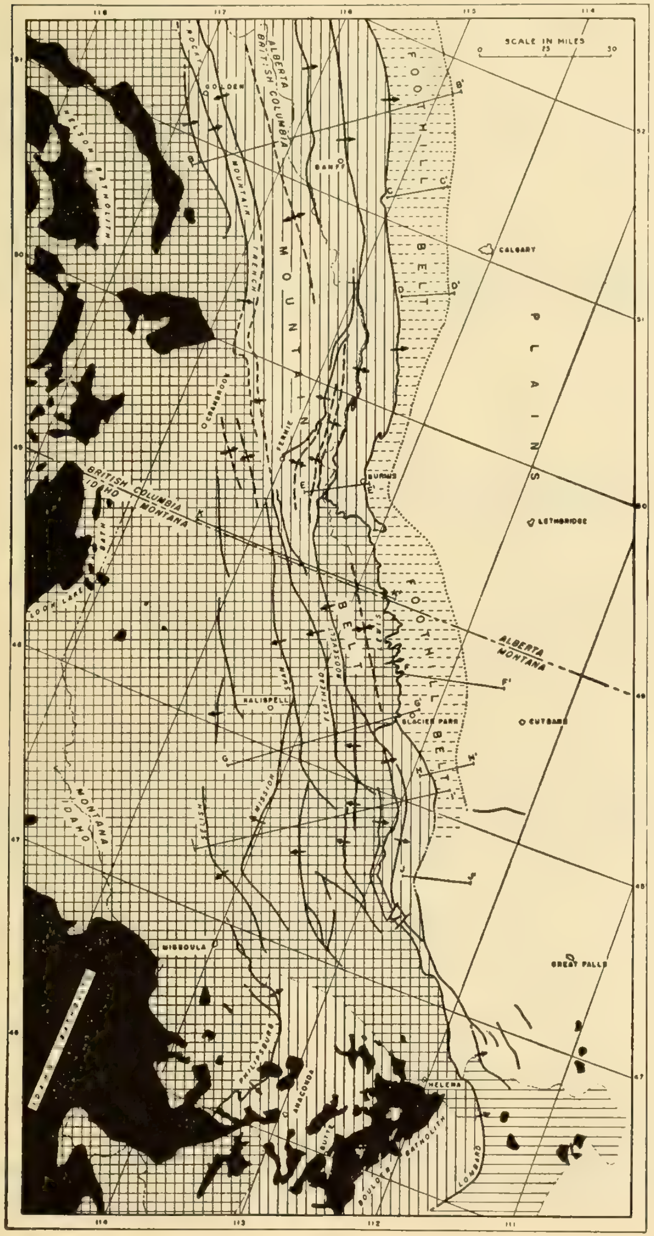
Richardson Mountains form a straight wall 175 miles long extending northerly from Peel River near longitude 136 degrees nearly to the Arctic coastal plain west of Mackenzie River delta. In the north they are more than 40 miles wide, and contain rugged, northerly trending asymmetrical ridges with peaks rising to heights of 5,000 feet or more. Throughout most of their length, however, they comprise a much narrower belt of steep-sided ridges, the flat tops of which lie mainly below 4,000 feet. No cirques or other evidence of alpine glaciation has been found in aerial photographs of even the highest peaks.

Mackenzie Mountains occupy a broad crescentic area, convex towards the northeast, stretching 425 miles southeasterly from south of Peel River near longitude 134 degrees nearly to Liard River at latitude 61 degrees. Their maximum width exceeds 100 miles. They are distinguished from Selwyn Mountains, which adjoin them on the southwest, not by any abrupt topographic boundary, but by absence of intrusions, conspicuous stratification, and more youthful topography. On the north and northeast they rise abruptly from the Mackenzie River lowland. In the main they comprise a compact mass of conspicuously layered, northwesterly trending ridges topped by peaks that commonly rise to elevations of more than 7,000 feet, and in some places are reported to exceed elevations of 9,000 to 10,000 feet. Small alpine glaciers are widespread. The Canyon Ranges, which form their northeastern front and occupy a belt up to 40 miles wide, include more subdued mountains and high plateau areas traversed by deeply incised river valleys.

Peel Plateau is a great triangular terrace occupying the angle between the east front of Richardson Mountains and the north front of Mackenzie Mountains. Its northeastern edge is in part a scarp rising 200 to 1,000 feet above the Plains region. The major rivers traversing the plateau, such as the Peel and Arctic Red, are deeply entrenched in the otherwise rather flat, glaciated upland surface.

Throughout most of their length Franklin Mountains lie a short distance east of and parallel with Mackenzie River. They extend from Fort Good Hope more than 400 miles southeasterly to the mouth of South Nahanni River and average less than 30 miles wide. They include, from north to south, Norman, Franklin, Camsell, and Nahanni Ranges, each comprising a number of parallel north to northwesterly trending ridges. In places they reach heights of 5,000 feet.

The Rocky Mountains form the eastern front of the Cordilleran region in British Columbia. Here they rise sharply from the comparatively flat Plains region, through a Foothills belt, to peaks reaching elevations of 10,000 to nearly 13,000 feet. These mountains, with their eastern foothills, have a maximum width of about 100 miles, and extend from the International Boundary at longitude 114 degrees 850 miles northwesterly to Liard River. At their northwest end, they are separated from Selwyn and Mackenzie Mountains by a distance of more than 100 miles. They have been carved from a thick series of sedimentary strata of rather simple structure, and the resulting layering, visible from great distances, at once distinguishes them from most other mountains in British Columbia. They consist of a series of overlapping ranges that trend northwest and, on the whole, have precipitous eastern faces and much less steep western slopes. Individual ranges are broken or terminated by deep cross-valleys, and the whole mountain mass is crossed by several deep depressions having comparatively low heights at the divides (Lord *et al.*, 1947).



DIVISIONS OF CANADIAN AND MONTANA ROCKIES

The two divisions of the Canadian and Montana Rockies generally recognized are the mountain belt and the foothill belt. The latter is commonly referred to simply as the foothills. The two divisions are shown on the map of Fig. 20.2. The western limit of the mountain belt of Laramide age in Canada is recognized by some as the remarkably regular depression called the Rocky Mountain trench. It extends from Liard River southeast for 800 miles to Flathead Valley in Montana. This serves as a convenient physiographic boundary of the Canadian Rockies, but as a structural boundary it is not secure. The Rocky Mountain trench may be traced southward to Kalispell in Montana, but from this point southward three great valleys exist, any one of which might be chosen for the trench. Clapp (1932) believes all the ranges of western Montana should be included in the Rocky Mountain system because they are alike structurally and stratigraphically. There, even more so than in Canada, the relation of the Nevadan and Laramide orogenic belts and the position of their boundary, if a common one, are little known and still speculative.

MOUNTAIN BELT

The mountain belt is made up of imposing front ranges such as the Lewis Range (Glacier Park) and the Canadian Rockies of the Banff and Jasper areas, as well as many large ranges to the west. All ranges trend approximately parallel with each other in a north-northwest direction, except from the Idaho batholith southward, where later deformation has imposed a topography discordantly in places across the Laramide structural trends. Compare Fig. 20.2 of this book with Raisz' *Landforms Map* (1939).

The mountain belt may be divided into two parts according to the

Fig. 20.2. Tectonic map of the Canadian and Montana Rockies showing their major divisions, the chief faults, the intrusions, and the lines of cross sections. Cross-ruled area is, with minor exceptions, folded and faulted Beltian (Proterozoic) rocks. Vertically ruled area is folded and faulted Paleozoic and Mesozoic rocks. The horizontally dashed area is the folded and faulted Mesozoic foothills belt. Horizontally ruled area is Beltian or underlain by Beltian but not part of the folded and thrust belt. Structures associated with the intrusions not shown. The arrows through the faults indicate the direction of movement of the overriding sheets.

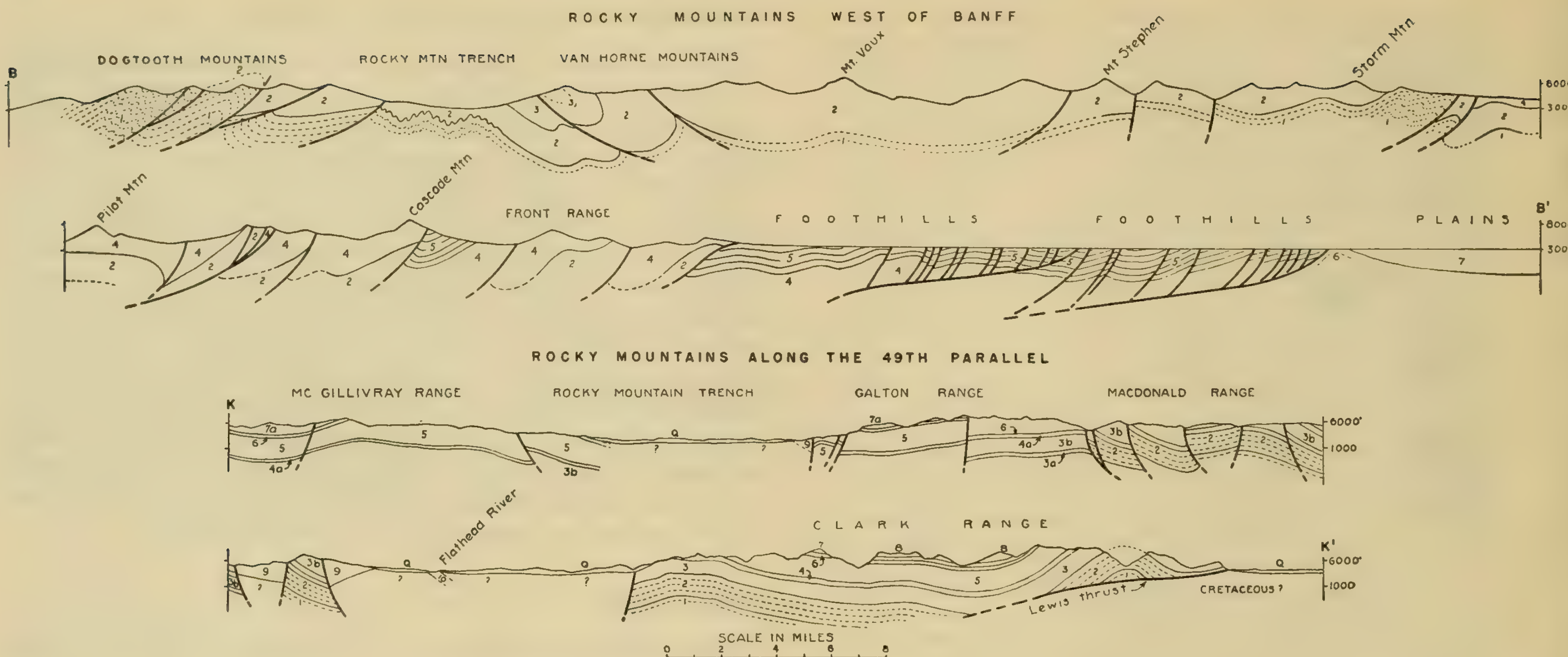


Fig. 20.3. Sections through the Canadian Rockies from the Rocky Mountain Trench to the Plains. Section B—B′ after Evans, 1932. 1, Beltian (?); 2, Cambrian; 3, Ordovician and Silurian; 4, Devonian and Carboniferous; 5, Mesozoic; 6, Edmonton (Montanan); 7, Paskapoo (late Paleocene). Section K–K′ after Daly, 1912. 1, Waterton dolomite; 2, Altyn limestone (1 and 2 Beltian); 3, Appekunny argillite; 3a, Hefty sandstone; 3b, MacDonald argillite; 4, Grinnell argillite; 4a, Wigwam sandstone and argillite (3, 3a, 3b, 4, and 4a Lower Cambrian); 5, Siyeh limestone; 6, Purcell lava; 7, Sheppard dolomite; 7a, Gateway argillite; 8, Kintla argillite; 8a, Phillips argillite and quartzite; 8b, Roosville argillite (5, 6, 7, 7a, 8, 8a, and 8b Middle Cambrian); 9, Mississippian and Devonian limestone; 10, Kishenehn clays (Miocene).

formations involved. One part is made up almost entirely of the formations of the great Beltian group, and the other of Paleozoic and Mesozoic formations. The two parts are designated by cross ruling (Beltian) and vertical ruling (Paleozoic and Mesozoic) on the map, Fig. 20.2. The Beltian division lies to the west except at the international boundary, where it extends to the east front of the mountain belt and adjoins the foothills belt, thus dividing the Paleozoic and Mesozoic division into a northern (Canadian) and southern (Montana) segment.

The stratigraphy and structure of the mountain belt are shown in a series of cross sections in Figs. 20.3 to 20.7. Section B–B′, K–K′, G–G′, and I–I′ are especially intended to typify the structures in southern Canada and in Montana.

A great thrust fault is the dominant feature along the eastern margin of the mountain belt. At the east base of the Lewis and Lewis and Clark ranges it is called the Lewis thrust and has been extended southward 150 miles in Montana to the Lombard thrust (Clapp, 1932) and northward from the international border at least 150 miles (*Calgary Sheet*, of the Canadian Geological Survey, Alberta, 1928). The great fault has several

branches as can be seen on the map, Fig. 20.2, and it is not obvious everywhere which should carry the name. It is a low-angle thrust. See sections G–G′, I–I′ and K–K′. In general, metamorphosed rocks of Beltian age have been thrust up and over shales and sandstones of Mesozoic age. The Lewis thrust in places is complex, consisting of several closely spaced parallel faults with considerable drag folding. In other places the fault is a single fracture, and the rocks on either side have not been greatly disturbed. For example, on the north side of Cut Bank Creek Valley, the Altyn limestone of Beltian age appears to rest almost conformably upon relatively uncrushed carbonaceous shale of the Colorado formation of middle Cretaceous age.

The fault in Glacier National Park and southern Alberta is very well known from the writings of Willis (1902), Campbell (1914), Daly (1912), Clapp (1932), and Billings (1938). Here the fault has a lower dip than

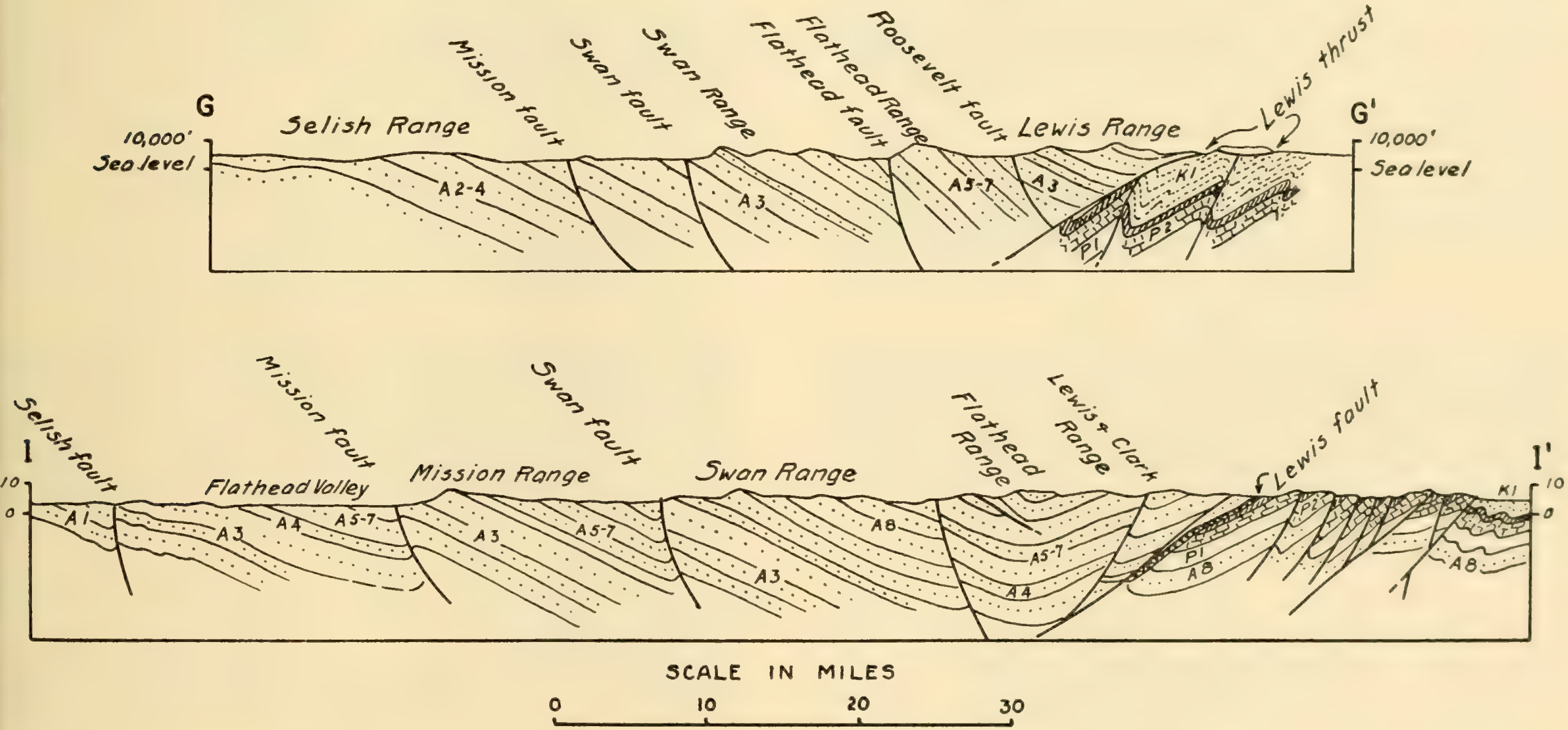


Fig. 20.4. Cross sections in northwestern Montana showing Lewis thrust and related structures. A1 to A8, the Beltian formations; A1, Prichard argillite; A2, Altyn siliceous limestone; A3, Appekunny quartzite and argillite; A4, Grinnell argillite; A5, Newland limestone and argillite; A6, Spokane argillite and quartzite; A7, Helena argillaceous limestone; A8, Missoula group, chiefly argillites, quartzites, and sandstones; P1, Lower Paleozoic formations; P2, Upper Paleozoic formations; K1, Lower, Middle and Upper Cretaceous. After Clapp, 1932.

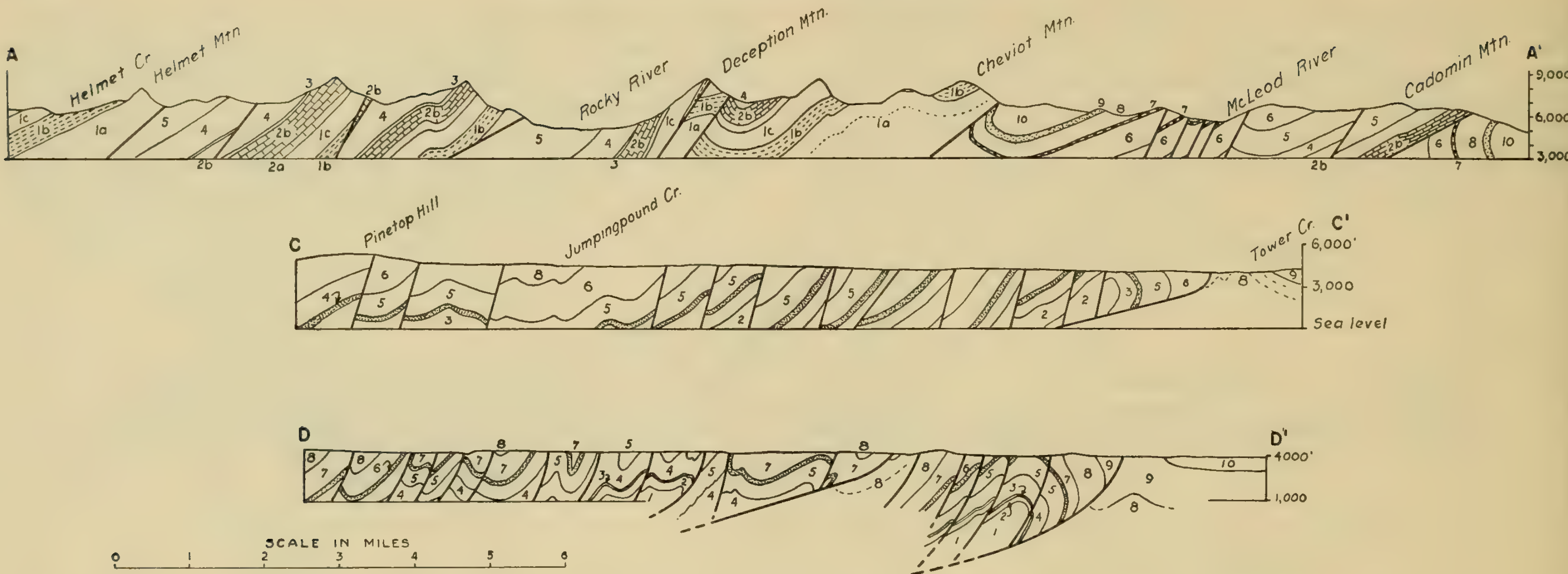


Fig. 20.5. A, section through Canadian Rockies at Mountain Park, Alberta, after MacKay, 1929. Section is north of limits of index map, Fig. 20.2. 1a, 1b, and 1c, Devonian; 2a, Banff shale (Mississippian); 2b, Rundle limestone (Mississippian ?); 3, Rocky Mountain quartzite (Pennsylvanian ?); 4, Spray River formation (Triassic); 5, Fernie shale (Jurassic); 6, Nikanassin shale and sandstone; 7, Cadomin conglomerate; 8, Luscar shale and sandstone; 9, Mountain Park sandstone, shale, and conglomerate (Lower Cretaceous); 10, Blackstone shale and sandstone (Colorado). The thrust under Cheviot Mtn. cuts the Big Horn and Wapiabi formation of Colorado age and the Brazean formation of Montana age, all younger than the Blackstone.

C, section near Jumpingsound Creek, Alberta, after Hume, 1932. 1, Kootenay sandstone and shale; 2, Blairmore sandstone and shale (1 and 2, Lower Cretaceous); 3, Lower Alberta shale; 4, Cardium sandstone and conglomerate; 5, Upper Alberta shale (3, 4, and 5. Colorado); 6, Belly River sandstone and shale; 7, Bearpaw shale; 8, Edmonton sandstone and shale (6 and 7, Montana; 8, Montana ?); 9, Paskapoo sandstone and shale (late Paleocene).

D, section through Turner Valley structure, Alberta, after Hume, 1931. 1, Paleozoic limestone; 2, Fernie shale (Jurassic); 3, Kootenay sandstone and shale; 4, Blairmore sandstone and shale (3 and 4, Lower Cretaceous); 5, Lower Alberta shale; 6, Cardium sandstone (5 and 6, Colorado); 7. Upper Alberta shale (Colorado and Montana); 8, Belly River shale and sandstone (Montana); 9, Edmonton sandstone and shale (Montana ?); 10, Paskapoo sandstone and shale (late Paleocene).

in most places—7 degrees—and the fault surface has either been warped or was uneven when formed. Two conspicuous klippen composed of Beltian rocks on Mesozoic shales are known as Chief Mountain and Divide Mountain, and near the headwaters of Ole Creek in the southern part of the park there is a window of Mesozoic shales entirely surrounded by Beltian rocks. South of Glacier National Park, the dip of the fault is generally steeper, but beyond Fiord Creek, 75 miles south of the park, the fault flattens out, and another window 3 miles long and half a mile wide appears. Southeastward, still, it becomes steeper, and eventually it is believed to join the Lombard thrust which has a dip of 40 degrees to the west and northwest.

Ross and Rezak (1959) conclude that the horizontal displacement of the sheet was at least 15 miles, probably 35 miles, and possibly more. They note the absence of erosional debris or an irregular land surface over which some geologists had suggested the sheet rode, and postulate the fault surface to be a shear.

Whereas some thrusts and thrust complexes clearly exhibit characteristics of gravity down-slope transport, it is difficult for the writer to con-

ceive of thrusts such as the Lewis to originate in any other way than by compression of a considerable thickness of the crust. The concept of compression is deeply ingrained in the literature of the Rocky Mountains, and the representations in this and following chapters reflect these views. They are challenged only if recent workers have taken a different view or if the writer feels strongly in favor of gravity induced movements.

West of the Lewis thrust and between the two tear faults that bound the Beltian segment is a broad syncline in the Beltian strata. See section I–I′, Fig. 20.4. West of the syncline, or on its west flank, a number of fairly high-angle faults that dip eastward repeat the formations (Clapp, 1932). The beds dip eastward at angles ranging from 20 to 50 degrees, and the faults dip more steeply than the beds, generally at angles of 60 to 80 degrees. These western faults, together with the eastern, form a set of huge downward-pointing wedges. Clapp estimates the amount of throw of the western high-angle thrusts to be from about 10,000 to 30,000 feet.

Each of the western faults follows closely the west base of one of the ranges which appear to have been uplifted along the faults. The faults have been named for their respective ranges (Clapp, 1932).

In addition to the eastern and western thrusts, there are steeply dipping transverse faults of both reverse and normal categories. They have displacements up to 10,000 feet. The transverse faults are most abundant in the southern part of the area north and northeast of the batholiths. One fault of a singular category has been mapped that parallels the beds but dips at a very low angle and is normal (Clapp, 1932).

Clapp relates the various groups of faults in the following way:

> It appears as if the forces causing faulting acted from the southwest, first uplifting and folding the rocks, then breaking them along the longitudinal (western) thrust faults. Later, the deformation continued to such an extent that relief from the stresses came by overthrusting (to the east). The two sets of transverse faults seem to be a still later effect of the continued pressure from the southwest, and consequent elongation to the northwest and southeast. . . . As the compressive forces acting from the southeast lessened, normal strike faults with low dip relieved the vertical pressures resulting from the great height of the uplifted rocks.

Normal faults, probably of late Cenozoic age, are also present and will be discussed under a later heading. Most of this faulting appears to have taken place along the much earlier longitudinal thrust faults.

In Canada, at least from Jasper at latitude 53° N southward to the border, the eastward and westward thrusts are found much in the same relations as in northwestern Montana. See section B–B′, Fig. 20.3. The strata instead of being mostly Beltian are mostly Cambrian, which in a broad way are synclinal, although a distinct and great anticline occurs west of Banff along the British Columbia–Alberta border.

A difference from the Montana division, however, is the nature of the western boundary. In Montana, the Beltian rocks continue westward under the entire terrane until the great intrusions make their appearance. In Canada, the belt of longitudinal thrust faults in Paleozoic formations is bounded on the west by the Rocky Mountain trench.

The belt of great imbricate thrust faults may be traced northward to Mountain Park (section A–A′, Fig. 20.5) and from there to ranges west of Fort Nelson. The Alaska Highway west of Fort Nelson, between miles 380 and 497, crosses two ranges, the Sentinel on the west and the Stone on the east. Here various rocks from Precambrian to Jurassic are exposed. The section, according to Laudon and Chronic (1949), is:

Triassic strata	
Black shale and black limestone	500 feet
Unconformity	
Mississippian strata	
Kindle formation: gray, silty limestone and chert	300–400
Unconformity	
Devonian strata	
Ft. Creek shale: black, pyritic shale	800
Ramparts limestone: massive, tan, gray, and black limestone	1500
Muncho limestone: gray and black limestone	600
Unconformity	
McConnell limestone: gray and black limestone	680
Unconformity	
Silurian strata, entirely Niagaran in age	
Roninng limestone: gray and black, cherty, dolomitic limestone	1200
Cambrian (?) strata	
MacDougal sandstone: tan sandstone	"thin"
Precambrian rocks	
Quartzite, slate, marble, and schist, intruded by basic igneous rocks	

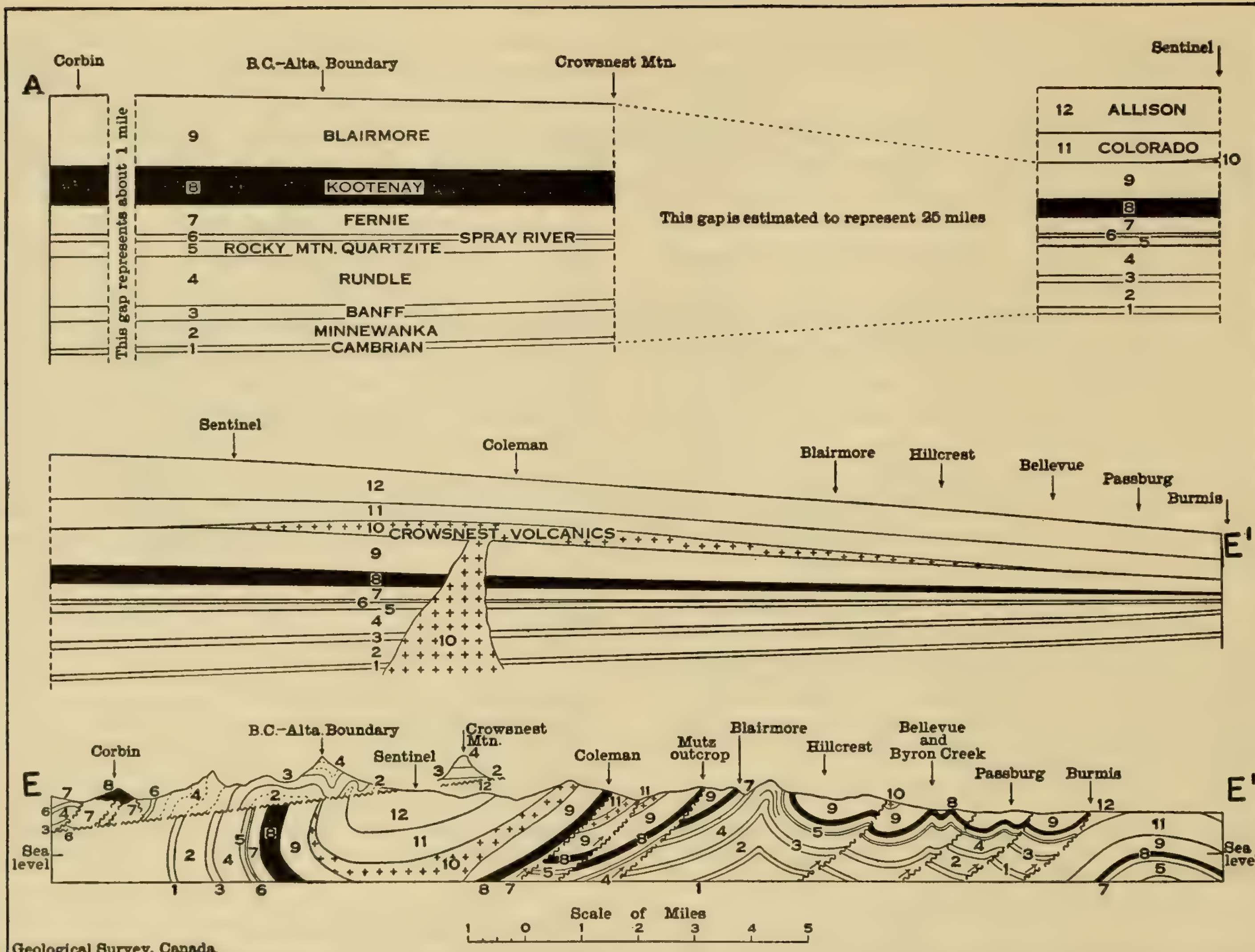


Fig. 20.6. Section E—E′ from Corbin to Burmis before and after the Laramide orogeny, after MacKay, 1932. (See Fig. 20.2 for the line of section.) 2, Devonian; 3 and 4, Mississippian; 5, Pennsylvanian; 6, Triassic; 7, Jurassic; 8 and 9, Lower Cretaceous; 10, 11, and 12, Upper Cretaceous.

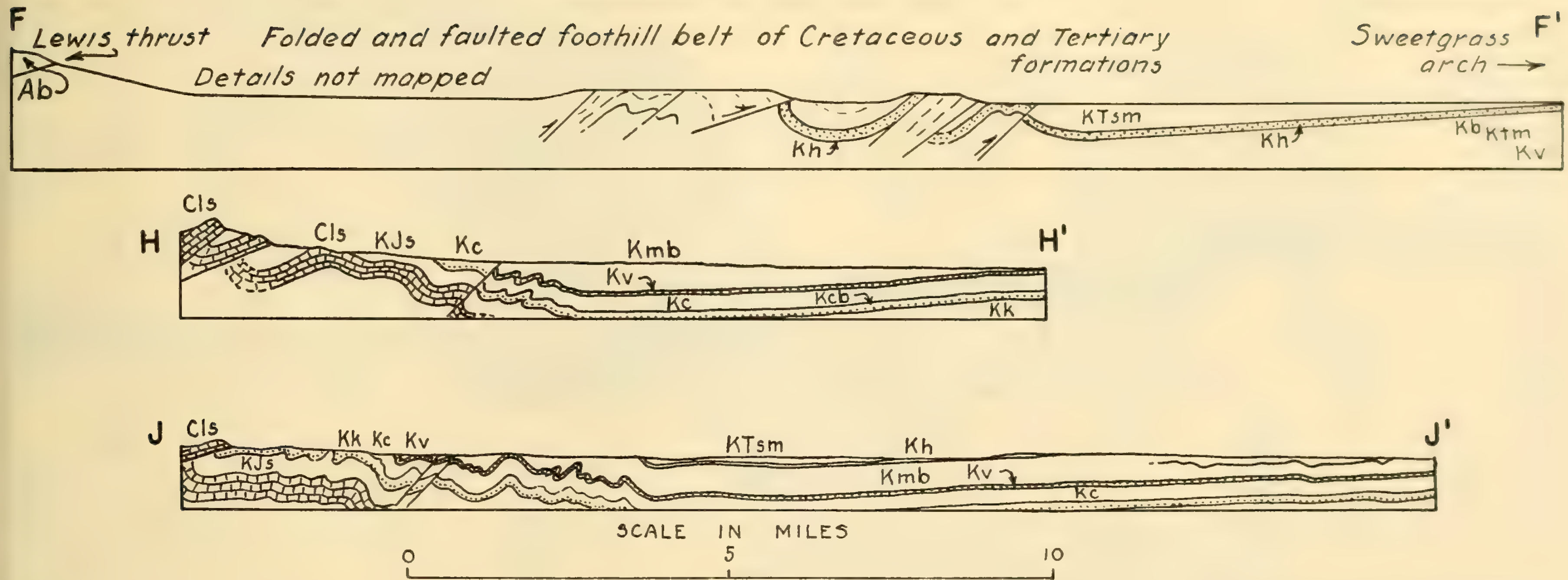


Fig. 20.7. Cross sections of Foothill structure in northwestern Montana. Upper section, F—F′, after Stebinger, 1916. Lower two sections, H—H′ and J—J′, after Stebinger, 1918. KTsm, St. Mary River formation; Kh, Horsethief sandstone; Kb, Bearpaw shale; Ktm, Two Medicine formation; Kmb, Bearpaw and Two Medicine undifferentiated; Kv, Virgelle sandstone; Kc and Keb, Colorado shale; KK, Kootenai formation; KJs shales and sandstones undifferentiated, belonging to Colorado, Kootenai, and Ellis (Upper Jurassic).

The structure of the two ranges is one of folding and thrust faulting typical of the Canadian Rockies farther south. The cross section of Fig. 20.10 illustrates the structure along the highway from miles 375 to 443.

See Figs. 37.1 and 39.14 and related text for brief discussion of the Mackenzie, Franklin, and Richardson Mountains in the far north.

FOOTHILL BELT

Sections B–B′, C–C′, and D–D′, Figs. 20.3 and 20.5, are typical of the folded and faulted foothill belt in Canada. Sections F–F′, H–H′, and G–G′, Figs. 20.4 and 20.7 are examples of the structure of the foothills in northwestern Montana. The belt ranges in width from 5 to 25 miles and extends from north central Montana (southwest of Cutbank, Fig. 20.2) northwestward to at least the 54th parallel, a distance of 500 miles or more. The foothills preserve remnants of early erosion surfaces, and are topographically low and related more to the Great Plains than to the mountains, but internally their structure is complex and reveals a great deal of compressional deformation. They are composed for the most part of the Cretaceous shales which have been easily eroded. The prevalence of the weak shales probably explains the reduction of the belt to one dominated by low, graded slopes. Only in the cores of a few anticlines are Paleozoic beds exposed (section H–H′, Fig. 20.7). The most common conception is that small reverse faults of a few hundred feet displacement are numerous, and that these terminate downward in major low-angle thrusts (Hume, 1926, 1931; Goodman, 1932). The anticlines and synclines that exist

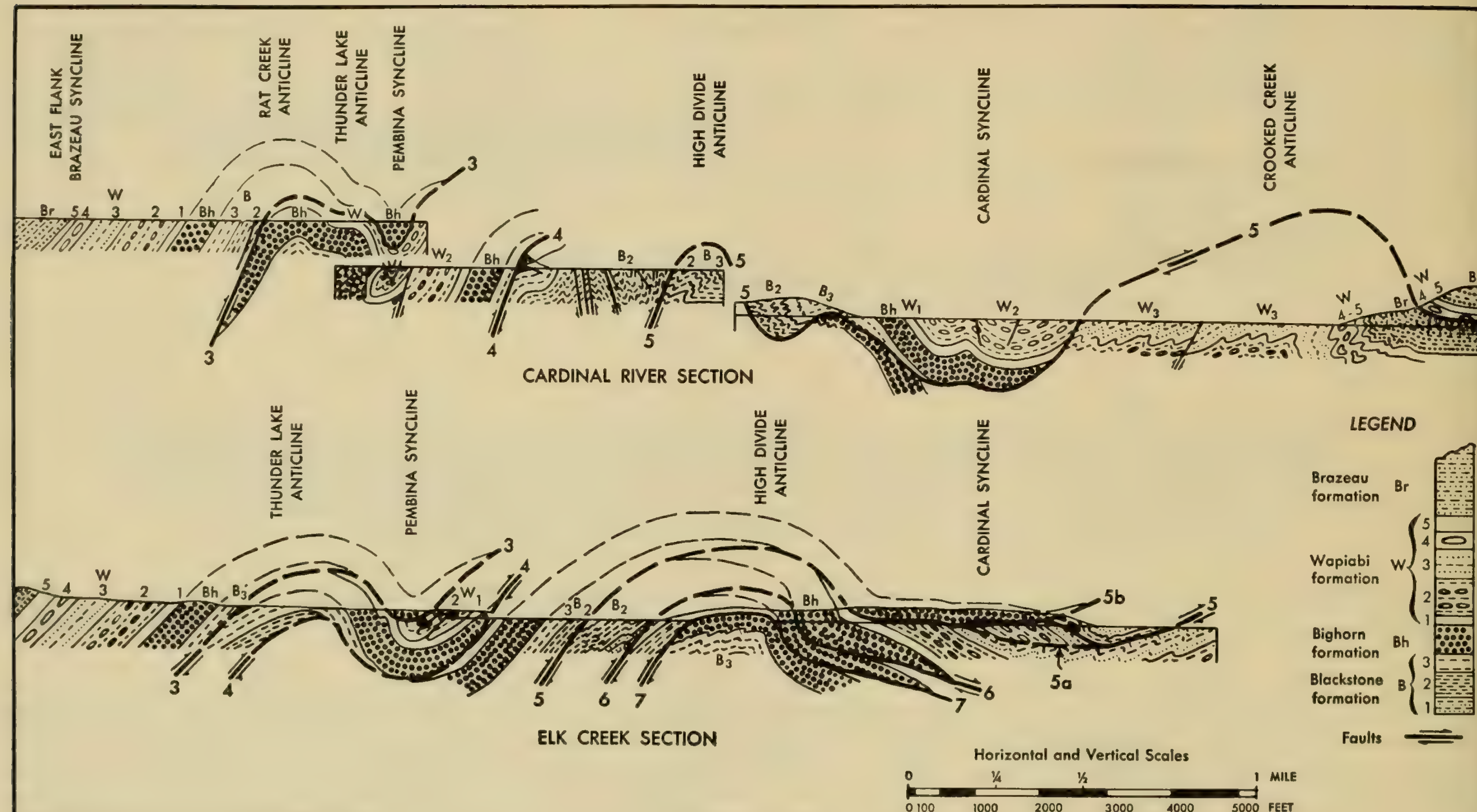


Fig. 20.8. Geologic sections along Elk Creek and Cardinal River in the southern part of the Cardinal district. After Hake *et al.*, 1942. See map, Fig. 20.9.

between the reverse faults are generally overturned toward the east and, in harmony with the faults, represent eastward movement of the thrust sheets.

In Montana, much of the foothill belt is covered with glacial drift, and the structures there, especially just south of the border, are not well known. See section F–F′, Fig. 20.7. Farther south, folding and overturning to the east seems to dominate reverse faulting (sections H–H′ and J–J′ Fig. 20.7).

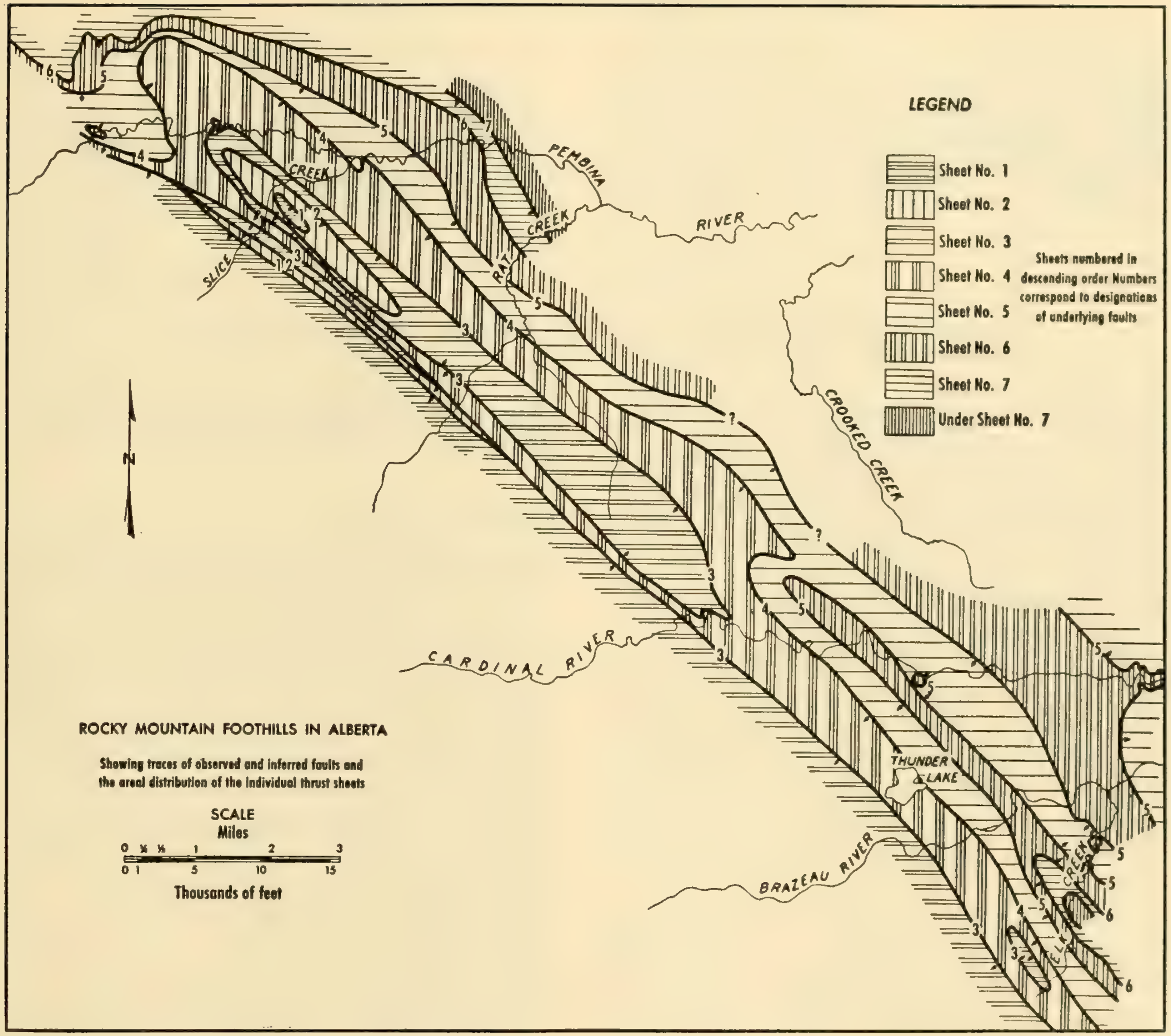


Fig. 20.9. Fault pattern of Cardinal district. After Hake *et al.*, 1942.

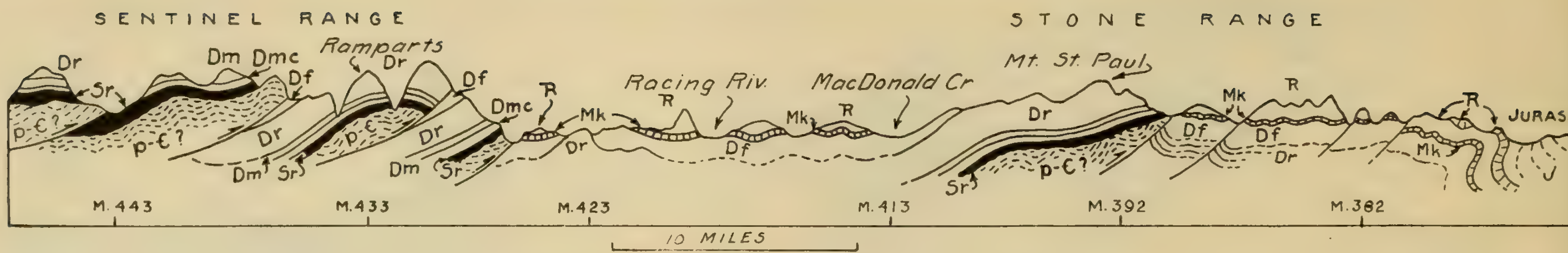


Fig. 20.10. Cross section along Alaska Highway, west of Fort Nelson, in northeastern British Columbia, between miles 375 and 443. After Laudon and Chronic, 1949.

North of Calgary, Alberta, and east of Jasper National Park in the Upper Brazeau River foothill area, Hake *et al.* (1942) have mapped a group of low-angle thrusts. The thrusts divide the Upper Cretaceous sediments into thin sheets that have been strongly folded. See sections of Fig. 20.8 and map, Fig. 20.9). The thrusts are considered noteworthy (1) because they are developed in weak beds, and the fault planes lie at an angle to the bedding so that the sheets themselves are not competent to have transmitted the thrust which caused the displacement; (2) because the faults bear an exceptionally systematic relation to the bedding. The investigators believe these faults developed in an asymmetrical syncline, and the faulting and the attendant crumpling relieved stresses which in other folds and in other stratigraphic sections are relieved by bedding-plane slippage. The thrusts are thought to be confined to the Mesozoic section and the major syncline from which they developed, and to die out completely with depth without producing any dislocation of the Paleozoic rocks.

It is evident that this concept is at variance with the more commonly portrayed one of many small high-angle reverse faults meeting a major low-angle thrust at depth, but the authors think that their theory may have widespread application in the foothill belt. This is confirmed by Scott (1954), who describes much the same structure as Hake *et al.* (1942) and repeats that it is found in other parts of the foothill belt besides the Brazeau and Cardium areas. He thinks that two distinct episodes of compression occurred, first thrusting of the thin sheets, and second, folding of the thrust sheets.

AGE OF THRUSTING

The Cretaceous and Paleocene formations of the Rockies, foothills, and plains of Alberta preserve the record of orogeny in the region to the west. See correlation charts, Figs. 20.11 and 20.12. Warren (1938) has summarized the evidence, and Fig. 20.13 is an attempt to show in diagram what he has said in words. Incorporated in the diagram are also Evans' ideas of the origin of the Rocky Mountain trench, and in addition, the concept of post-Laramide graben-type faulting.

The Kootenay and Blairmore of Early Cretaceous age thicken westward, and conglomerates become abundant. A basal conglomerate of the Blairmore is believed to represent the first pronounced uplift to the west. The formations are exposed in the Canadian Rockies in Elk River at Crowsnest Pass (see Fig. 20.8), and some of the pebbles and boulders of the conglomerates are medium- to fine-grained granite and granite porphyry that could only come from the Selkirks (Evans, 1932). This seems adequate evidence to date the first uplift of the Selkirks and to indicate that the Rockies had not yet come into existence but were a site of deposition. Since the Blairmore is Aptian and Albian, it is evident that the Selkirks first rose in latest Jurassic or earliest Cretaceous time. The Lower

Cretaceous sediments have been charted for the entire western part of the continent on the tectonic map, Plate 11.

The Kootenay and Blairmore formations are continental in origin, and reflect an uplift of the region to the west and a source of abundant feldspathic sediments. The next formation, the Colorado or Alberta shale, is marine and represents a marine invasion. See chart, Fig. 20.11. The Belly River that followed the Colorado is continental and resembles the Blairmore. It reflects renewed uplift on the west. A local sea invaded the southern foothills belt from the south, and in it the Bearpaw shale was deposited. A continuation of uplift in the Selkirks resulted in the deposition of the continental Edmonton. Then a period of erosion occurred that represents the Lance, early Paleocene, and middle Paleocene (Russell, 1932); and following it, the upper Paleocene Paskapoo sandstone and shale were deposited. In the foothills and plains no angularity between the Edmonton and Paskapoo has been noted.

All the foregoing Cretaceous formations and also the uppermost Paleocene beds are folded and faulted in the foothills (Russell, 1932), and therefore the main deformation of the frontal Canadian Rockies occurred in post-Paleocene time.

A second but milder orogeny is noted by Bostock *et al.* (1957):

> In the Flathead Valley, west of Clark Range, the Kishenehn formation of very late Eocene or very early Oligocene (Upper Duchesnean) age unconformably overlies early Mesozoic strata which are involved in the structures of the southern Rocky Mountains. The strata are gently folded, dipping mainly about 30 degrees northeast. These observations indicate two phases of deformation, the first, pre-Kishenehn and post-early Mesozoic, probably post-Paleocene in age, during which the main orogenic movements took place, and the second, post-Kishenehn in age, during which the Kishenehn beds were tilted. Conglomerates of the Kishenehn carry pebbles of Proterozoic rocks indicating that these rocks were exposed to adjacent ranges following the first phase of the deformation.
>
> Further evidence on the age of the uplifts associated with the orogenic movements in the southern Rocky Mountains is found in conglomerates on the Plains of southern Saskatchewan. The products of erosion from the uplift of the southern Rocky Mountains during two phases of the deformation are thought to be represented by gravels in the Cypress Hills region which carry pebbles of the distinctive Proterozoic rocks and bracket in age the time of deposition of the Kishenehn formation. Mammalian fossils in the Swift Current Creek beds are Uintan (late Eocene) and those in the Cypress Hills formation are of Chadronian (early Oligocene) age. In summary, the first and main deformation in the southern Rocky Mountains took place in the interval between the Paleocene and the late Eocene. Uplift and erosion occurred in the late Eocene (Uintan), followed by relative quiescence during the deposition of

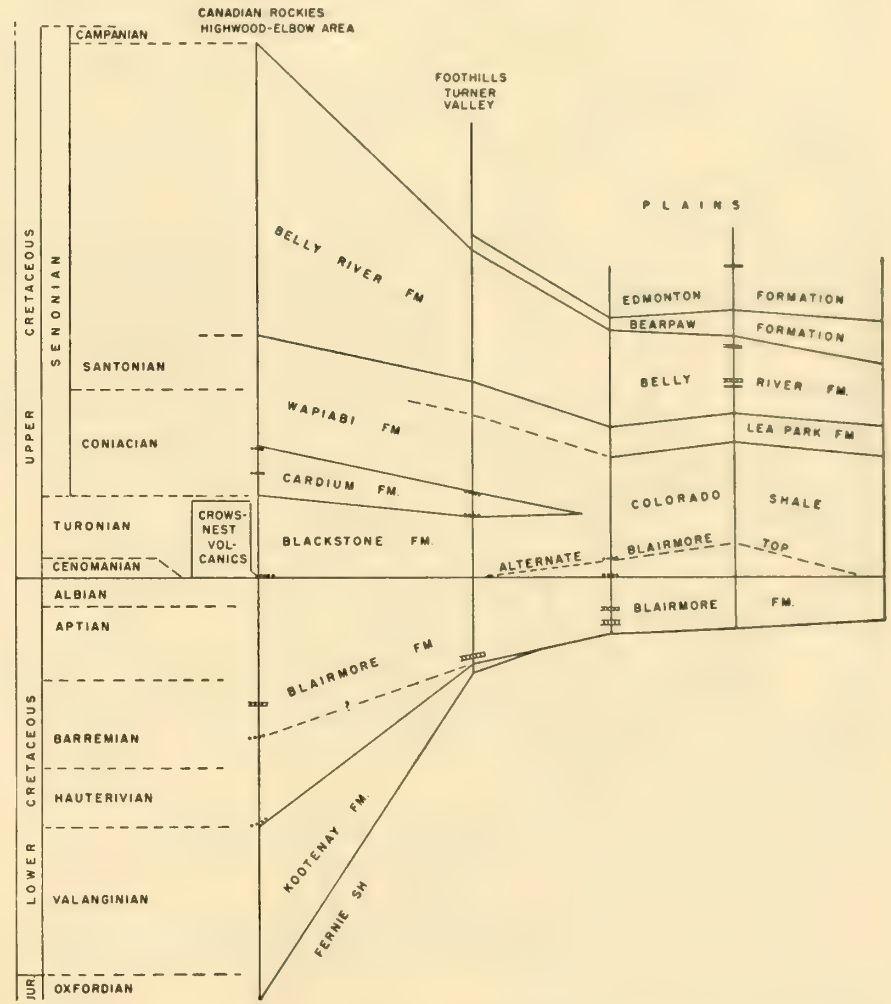


Fig. 20.11. Cretaceous formations south of Calgary, Alberta. Lithology is sandstone and shale except where conglomerates and limestones are indicated. The Colorado and Bearpaw sediments are marine, the rest brackish and fresh water. After Thompson and Axford (1953).

		WEST SIDE OF ALBERTA SYNCLINE			EAST SIDE OF ALBERTA SYNCLINE			
	WESTERN UNITED STATES	SOUTHWEST AREA		NORTHWEST AREA	SOUTHEAST AREA	NORTHEAST AREA		
		OLDMAN, CASTLE RIVERS, ETC.	LANGFORD CREEK MAP-AREA	HIGHWOOD AND BOW RIVERS	OLDMAN RIVER	LITTLE BOW RIVER	BOW RIVER	RED DEER RIVER
PALEOCENE	FORT UNION	PORCUPINE HILLS	PORCUPINE HILLS	PASKAPOO	PORCUPINE HILLS	PASKAPOO	PASKAPOO	PASKAPOO
		UPPER WILLOW CREEK ZONE 'E'			UPPER WILLOW CREEK			
UPPER CRETACEOUS	LANCE	LOWER WILLOW CREEK 'D' 'C' 'B' 'A'	LOWER WILLOW CREEK	EDMONTON	LOWER WILLOW CREEK	WILLOW CREEK FACIES		UPPER EDMONTON
	FOX HILLS	ST. MARY RIVER	ST. MARY RIVER		KNEEHILLS TUFF / BATTLE EQUIVALENT / WHITEMUD EQUIVALENT	EDMONTON: ST. MARY RIVER FACIES	KNEEHILLS TUFF / BATTLE EQUIVALENT / WHITEMUD EQUIVALENT	KNEEHILLS TUFF / BATTLE EQUIVALENT / WHITEMUD EQUIVALENT
	PIERRE	BASAL MEMBER / BLOOD RESERVE			ST. MARY RIVER / BASAL MEMBER / BLOOD RESERVE	EDMONTON FACIES	LOWER EDMONTON	LOWER EDMONTON
	GSC	BEARPAW	BEARPAW		BEARPAW	BEARPAW	BEARPAW	BEARPAW

Fig. 20.12. Correlation of uppermost Cretaceous and Paleocene formations of southwestern Alberta. Reproduced from Tozer (1953).

the Kishenehn beds Duchesnean. Moderate deformation and renewed uplift took place in early Oligocene (Chadronian) time.

The Eocene age of the Lewis thrusting has been demonstrated fairly well by MacKenzie (1922) from an Eocene formation in the Flathead Valley, back of the Clark Range. Alden (1932) also believes the Lewis thrust occurred in Eocene time.

Evans (1932) presents two arguments to support an earlier age of the Selkirk system, west of the trench. The great Rocky Mountain trench and the structures of the Rocky Mountains trend parallel with each other, but the mountains west of the trench, viz., the Purcell Range, the Selkirk Range, and the mountains west of Columbia Lakes, trend nearly north at an acute angle to the trench, and are truncated by it. From these relations it seems that the trench is associated with the building of the Rockies which were formed later than the Selkirks.

Secondly, the Selkirk system contains many great intrusions; the Rockies only a few smaller ones. See Fig. 37.1. The intrusions have been related to the Coast Range batholith, and considerable evidence in Chapters 21 and 37 has been summarized that shows they are probably of Early or Mid-Cretaceous age.

The structural discordance and the great intrusions of the Selkirk system fit into the sedimentary record very well, and all three together demonstrate a fairly substantial case for the Early Cretaceous age of the

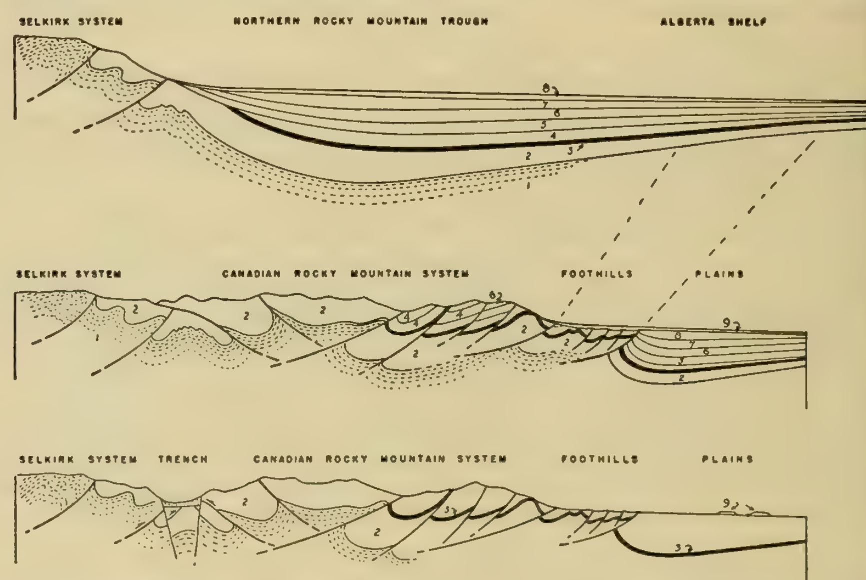


Fig. 20.13. Evolution of the eastern margin of the Selkirks and the Canadian Rockies. Idealized sections incorporating parts of sections B—B′ and E—E′. 1, Beltian; 2, Paleozoic, Triassic, and Jurassic; 3, Kootenay; 4, Blairmore; 5, Colorado; 6, Belly River; 7, Edmonton; 8, Paskapoo; 9, Lower Oligocene conglomerate. Upper diagram, growth of Selkirks during Cretaceous time and subsidence of northern Rock Mountain trough. Middle diagram, Laramide orogeny during the Eocene and the deposition of the Lower Oligocene conglomerate. Lower diagram, erosion of Rocky Mountain trench and Lower Oligocene conglomerate producing present aspect.

mountains west of the trench and an Eocene age for the Rockies east of it. The Rocky Mountain trench is probably still younger and of mid- or late Cenozoic age.

THE ROCKY MOUNTAIN TRENCH

In British Columbia. As previously indicated, a deep, wide valley separates the opposing Canadian Rockies on the east from the Selkirk system of ranges on the west in southern British Columbia. The Dogtooth, Purcell, and McGillivray ranges (see sections B–B′ and K–K′, Fig. 20.3) are parts of the Selkirk system that flanks the valley on the west, and the Van Horn, Brisco, and Galton ranges are examples of the Rocky Mountain system on the east. The great valley is so regular and continuous that it was called the Rocky Mountain trench by Daly.

It does not have a continuous downhill gradient, but within the trench are low divides that separate courses of several great rivers. The Ketchika River drains the trench northward from latitude 58° into the Liard River. South of latitude 58°, the Finlay River drains the trench into the Peace River which flows eastward through great canyons in the Rockies. The Parsnip River is a tributary of the Peace that extends southward nearly to the 54th parallel. The Frazer River occupies the trench from 54 to 53 N. Lat., and then the Columbia and its tributaries flow in the trench nearby to the international border.

Except for about 60 miles between the big bend of the Frazer River and latitude 55° the trench is sharply or fairly sharply defined from Kalispell, Montana, to beyond latitude 58°, nearly to the Yukon border, a distance of over 900 miles.

The Rocky Mountain trench lies at the boundary approximately between the Nevadan and Laramide orogenic belts. According to Bostock *et al.* (1957):

> Throughout most of its length it forms the approximate boundary between intensely deformed, altered, and intruded rocks characteristic of the western Cordillera, and the moderately deformed and comparatively unmetamorphosed strata that typify the eastern Cordillera. However, the trench does not everywhere coincide with this geological boundary; in several places it obliquely transects structures on both sides and, south of about latitude 50 degrees, the geological boundary lies east of the trench. North of this latitude the trench is known in several places to be the locus of extensive faulting and may have

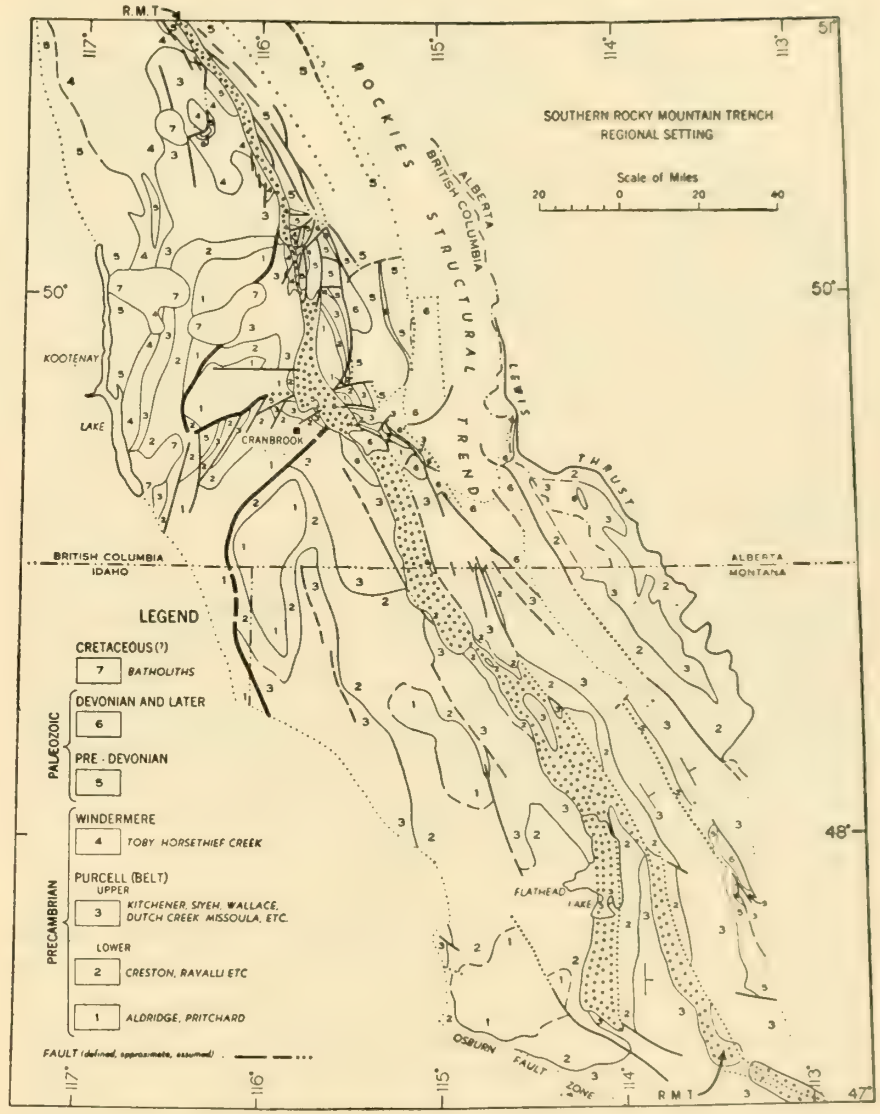


Fig. 20.14. Geology on either side of the southern part of the Rocky Mountain trench. Reproduced from Leech, 1959. Stippled zone is trench.

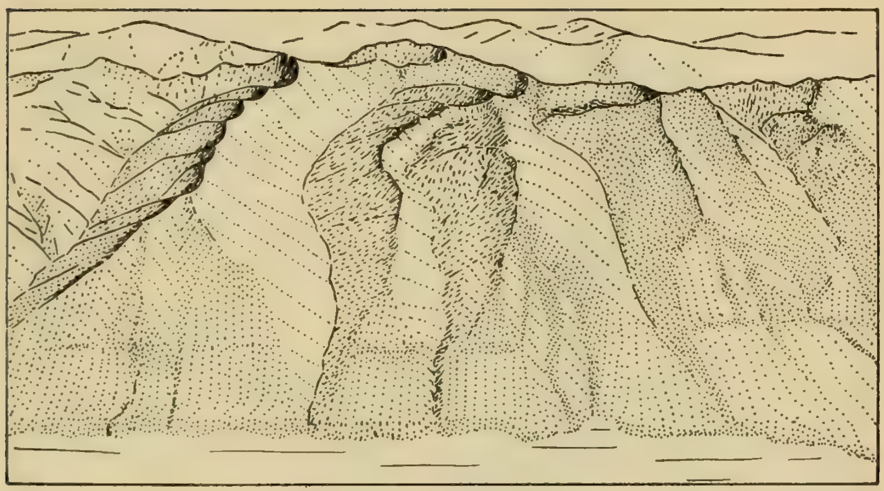


Fig. 20.15. Sketch of east face of Rocky Mountain trench at latitude 49°08′. Made from photograph by Leech, 1959.

been an active feature since the oldest Cordilleran disturbances. From latitude 50° to 51° 30′ or beyond, several longitudinal faults pass into the trench at small angles. Westerly dipping thrust faults cut the rocks of the Dogtooth Mountains into slices, and such a fault or fault zone is assumed to underlie the floor of the trench for many miles. Easterly dipping faults east of this part of the trench have been interpreted as underthrusts. Some long straight steep faults, such as the Redwall, may be dominantly strike-slip faults. North of Finlay Forks, rocks of the Sifton formation that floor the trench have been tilted and cut into long narrow slices by closely spaced faults that strike parallel with the trench. The fault slices transect, at a small angle, the strike of the Sifton strata and that of structures immediately east of the trench.

The long trench has been little studied until recently when a report on its nature immediately north and south of the international border has appeared (Leech, 1959). Previously it had been postulated to be the result of erosion following Cretaceous thrusting and folding (section B–B′, Fig. 20.3), or to be due to normal downfaulting, either of late Laramide or late Cenozoic age (Section K–K′, Fig. 20.3). In the first edition of this book it was postulated to be due to late Cenozoic graben-type faulting or rifting, and part of a great belt that extends from southwestern Utah to the Yukon.

Figure 20.14 is a map reproduced from Leech (1959) which shows the complex Laramide and Nevadan (?) structures on either side of the trench. The following points are mostly by Leech:

1. The trench in the Cranbrook area is particularly sinuous in contrast to its linear extent farther north.
2. It is asymmetrical, with the east flank high and of youthful fault-scarp topography (Fig. 20.15).
3. It contains outcrops of Paleozoic and Belt strata on its floor and does not appear to be as heavily alluviated as are some of the trenches further south in the United States.
4. It is probably of block fault origin but bounding normal faults are not everywhere apparent, especially in the sinuous section.
5. The postulated bounding normal faults are commonly disposed acutely to the older thrust faults.
6. Since the same formations appear on either side of the trench in this southern region, it is evident that here the rift is not an exact boundary between the Nevadan orogenic province on the west and the Laramide on the east.

In the Yukon. The trench loses its identity north of latitude 59°. The division between Nevadan and Laramide provinces also is difficult to identify, but probably swings northerly to lie east of the Selwyn Mountains.

About 100 miles northwest of Watson Lake on the Yukon–British Columbia border a remarkably strait valley, the Tintina, extends for about 400 miles northwesterly to the Alaskan border. See Fig. 39.1. Although not connected with the Rocky Mountain Trench it is in alignment with it, and a Tertiary filled valley at Watson Lake helps bridge the gap. According to Bostock *et al.* (1957);

> Major faults mark the course of the valley near Ross River Post and in the Glenlyon and southwest Mayo area, and major geological boundaries coincide with it in other places. Early Tertiary beds, only gently warped, outcrop at intervals along the valley floor, proving its early development as a physiographic feature.
>
> Shakwak Valley, another long straight lineament, extends from the Alaska boundary southeast through Kluane Lake almost to latitude 60 degrees. Through most of its length it forms a major geological boundary and is believed to mark a great fault zone. Evidence of recent movement is found in unconsolidated deposits along the valley floor. Southwest of Shakwak Valley in the Kluane area, a zone of overthrust faults is believed to form, with the Shakwak Valley fault, a graben structure enclosing upper Paleozoic to Tertiary rocks.

21.

IDAHO BATHOLITH AND THE OSBURN FAULT ZONE

EXTENT

The Idaho batholith extends from the vicinity of Boise northward through the center of Idaho into Montana, and has an area of over 16,000 square miles. Plutons of batholithic dimensions occur in the narrow northern end of Idaho, as if to link the main Idaho batholith to the Loon Lake, Colville, and Nelson batholiths. See Fig. 17.13. Smaller batholiths and stocks also occur nearby in western Montana, namely, the Boulder batholith near Butte, the Philipsburg stocks near Philipsburg, and other unstudied and unnamed batholiths to the south and southeast.

Most of the western, all of the southern, and part of the eastern borders of the batholith are blanketed by Tertiary strata, mainly Miocene volcanic rocks, and it is generally recognized that the batholith is much larger than that exposed and shown on maps.

COMPOSITION

Composition of Main Mass

According to Ross (1928), the Idaho batholith is composed mainly of quartz monzonite, although marginal facies are commonly granodiorite. In northern Idaho north from Pend Oreille Lake large plutons are composed of granodiorite, quartz-monzonite, and granite, and are regarded sufficiently similar to the Idaho batholith to be connected with it genetically and to bridge it to the Nelson batholith in British Columbia (Ross, 1928).

Later Anderson (1942) found the marginal facies to have been diorite originally, with only minor amounts of quartz, and that it was subsequently altered by widespread rising solutions rich in silica. Much quartz was added, and generally also smaller amounts of potash, feldspar, biotite, and sphene. Where little or no potash feldspar was added, as along the northwest margin of the batholith, a quartz-rich diorite (tonalite) was produced; where considerable amounts of potash feldspar were added, as along the south and southwest margin, granodiorite formed. The inner facies, upon consolidation, was less calcic than the marginal, and originally ranged from a diorite to granodiorite, with oligoclase rather than andesine. Postconsolidation emanations added considerable silica and potash and increased slightly the amount of biotite. Considerable added potash feldspar changed the rock to a quartz monzonite; less, to a granodiorite. Locally, enough potash was added to form granite; in places, a muscovite granite.

Younger Intrusives

The great batholith is now known to be composite and to contain plutons younger than the main mass of quartz monzonite or granodiorite. See the *Geologic Map of Idaho* by Ross and Forrester, 1947. Ross (1935) has described an intrusion in the Casto district that cuts the Laramide struc-

tures and perhaps even Miocene (?) beds. It consists of a pink granite to a quartz monzonite. The map of Fig. 21.2 shows this intrusion, as well as others of similar age and relation to the main batholith. The data were taken from the *Tectonic Map of the United States.* Ross also mentions pink granites in the northwest corner of Idaho and in British Columbia that probably cut Miocene (?) strata, and which he believes are distinctly younger than the Nelson batholith. These pink granites lead the writer to think of a pink granite in southwestern Montana which proved to be Precambrian in a thrust sheet which was later cut and displaced against Miocene (?) basin beds by a high-angle fault, as if in intrusive contact with them.

Anderson (1948) describes two areas of younger intrusives within the main batholith and says there are "many others." The younger intrusives are of two sets, one believed to have been emplaced at the close of the Laramide orogeny and the other in mid-Tertiary time. The early Tertiary magma was chiefly noritic, and the mid-Tertiary injections range in composition from dacite to rhyolite, with quartz monzonite porphyry and rhyolite porphyry most abundant. The Tertiary plutons within the main batholith are small and elongated. One, however, Anderson describes as 8 miles long and $\frac{1}{10}$ to $1\frac{1}{4}$ miles wide. They invade fault and shear zones, the main ones of which extend in a northeast direction.

Again in 1952 Anderson cites evidence that discrete masses of the granitic rock were emplaced under deep-seated conditions and others at much shallower depths. The deep-seated plutons include one that evolved while the major orogeny was taking place, and another which came in during the later, less intense stages of deformation. The shallower intrusions are those of Laramide and later age.

Border Zones

The Thatuna pluton, a satellite on the west, is principally granodiorite but grades into adamellite, tonalite, and granite. The Beltian strata which the Thatuna batholith intrudes are variably affected. In extremely fine-grained types, the contact is sharp and follows joint planes; but the contact with the granular quartzite is gradational through several hundred yards. By increase in feldspar the quartzite grades into igneous rock. To the southeast of the Thatuna pluton, thin layers of pegmatite and aplite are interlayered with paragneiss and diopside quartzite to form an extensive mass of gneiss. The belt of gneiss is 12 miles wide in Latah County, and extends for 15 miles at least into Clearwater County, where it borders the Idaho batholith. An extension of the Idaho batholith is believed to underlie the metamorphic belt (Tullis, 1944).

The Bitterrot Range of Idaho and Montana is largely a zone of gneiss and schist that borders the Idaho batholith on the northeast corner. It is a migmatite of the intrusion, according to Langton (1935), but according to Sydney Groff of the Montana Bureau of Mines and Geology (personal communication) it is a Precambrian terrane.

AGE

Consanguinity

The age of the great Idaho batholith is an important problem in the tectonic setting and, at the same time, a matter of controversy. The problem seems to be resolved into an issue between a Nevadan and a Laramide age.

The principal argument advanced for the Nevadan age of the Idaho batholith is its lithologic similarity to the batholiths of the Nevadan orogeny, specifically to the Nelson batholith (Ross, 1928). Since all the batholiths exhibit many variations in the granitoid series, generally, from diorite to granite, it does not seem possible to correlate them closely in age on the basis of lithologic similarity. It must be granted, however, that the granitoid character, together with great size and clustered grouping, seems to relate them to a common great orogenic belt and batholithic cycle. Nothing similar to the Idaho batholith occurs elsewhere in the Laramide orogenic belt.

Intrusive Relations

Near its southeastern end the batholith intrudes a thick series of Paleozoic strata. In this vicinity, isolated granitic masses similar and probably satellitic to the batholith cut Paleozoic strata as young as Pennsylvanian. Farther north the bordering formations are mostly Proterozoic (Beltian)

quartzites and slates. Still farther northeast in Montana the batholith is believed to be bordered chiefly by Beltian rocks, although little is known geologically of this region. On its west side, the bordering formations, in addition to the extensive Tertiary volcanics of later origin, include pre-Tertiary sedimentary and volcanic strata which are intruded by it. The pre-Tertiary rocks along the part of the western boundary north of Salmon River are mostly so metamorphosed as to make correlation doubtful, but along Snake River there are considerable thicknesses of Permian strata and some Triassic beds, both of which include volcanics. Small granite masses, presumably satellites, cut the Permian strata. The Thatuna batholith is one of these (see Fig. 17.13).

In numerous places, Tertiary strata, mainly Miocene (?) volcanic rocks, rest on the eroded surface of the batholith; and it is clear that much of the eastern part of the batholith now exposed was laid bare by erosion prior to the Tertiary volcanism. Some of the volcanic flows resting on the batholith may be as old as Oligocene (Ross, 1928).

Satellites (?)

The numerous plutons east of the Idaho batholith in western Montana, such as the Philipsburg (Calkins, 1915), Boulder (Knopf, 1913), and Marysville (Barrell, 1907), intrude either Cretaceous formations or older Mesozoic formations that were folded and thrust following the deposition of the Upper Cretaceous beds. The intrusions are distinctly discordant with the folds and thrusts and, as far as known, were all emplaced after the Laramide thrusting. They constitute a middle or late phase of the Laramide orogeny.

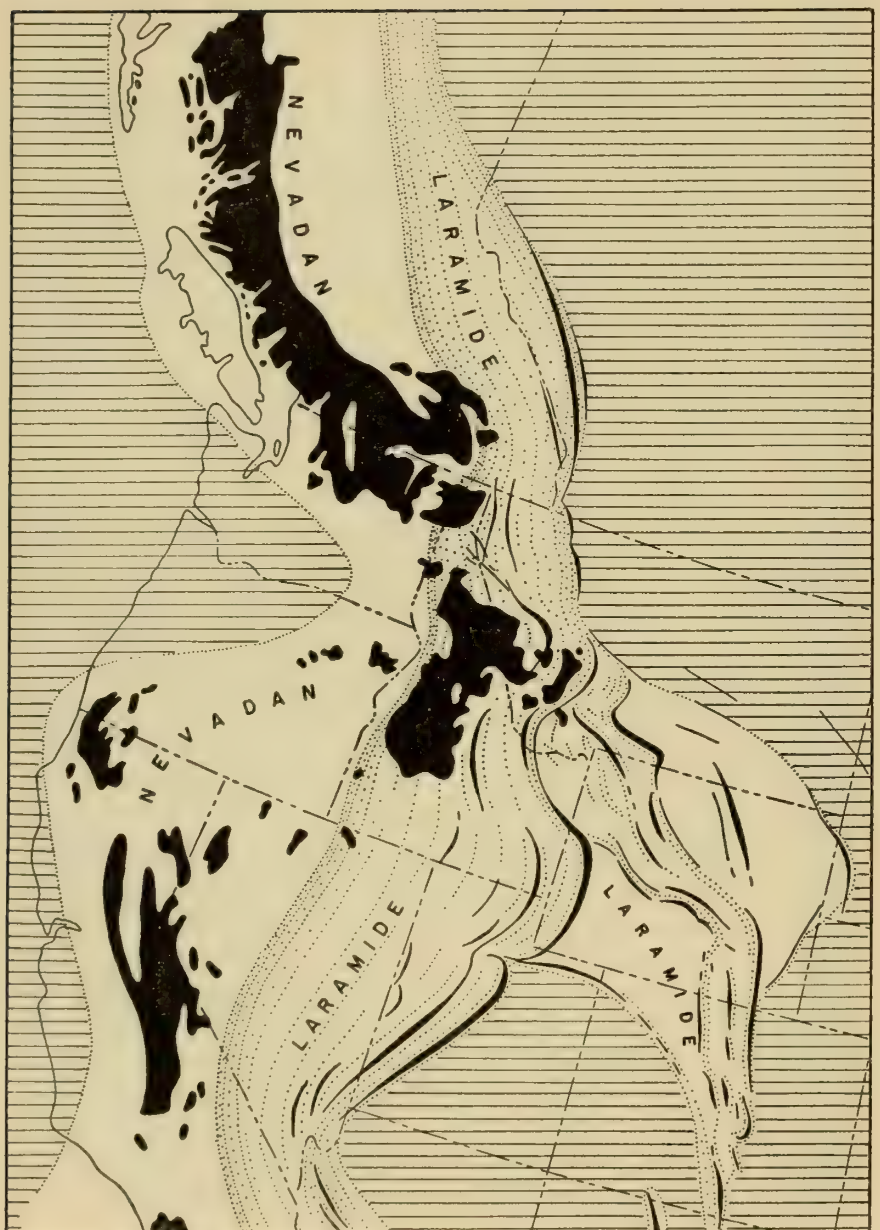
If the assumption is correct that the main batholith was intruded at the same time as its smaller eastern neighbors, then the great pluton must be Laramide in age and not Early Cretaceous or Late Jurassic (Nevadan). In further consideration of this line of evidence, it may be seen (Fig. 21.2 and *Tectonic Map of the United States*, 1944) that the Philipsburg thrust is truncated by the main eastward-extending appendage of the Idaho batholith. But this appendage is represented on the new *Tectonic Map of the United States* as a separate intrusion of later age than the main igneous mass. The representation comes of necessity when the main mass is shown as Nevadan. Details are not known, because the appendage has not yet been described in print.

The Casto intrusion is exposed along the axis of a broad anticline and involves both Permian (?) and Miocene (?) strata (Ross, 1935). Injections of pink granite into the Miocene (?) beds indicate the age of the pluton to be Miocene (?), according to Ross; but then, the exact age of the Tertiary beds is not known. Ross mentions other pink granites in the northwest corner of Idaho and in British Columbia that probably cut Miocene (?) strata and are distinctly younger than the Nelson batholith. The Nelson is believed to be earlier than Late Cretaceous because pebbles of its granite are found in the Blairmore conglomerate of Late Cretaceous age. The pink granites appear to be the youngest of the plutons, even considerably younger than the Boulder batholith (Ross, 1928).

Setting in Laramide Tectonic Plan

Figure 21.1 has been prepared to show in a broad way the relation of the Idaho batholith to the Nevadan and Laramide orogenic belts. In brief, the batholith is located at the junction of two arcuate segments of the Laramide belt, one extending from Canada into Montana on the north, and the other extending from Utah through Wyoming and southwestern Idaho on the south. A third major structural element, the zone of thrusting of the shelf ranges, converges here also. The converging of the three large elements of the Laramide orogeny at about the position of the Idaho batholith may be genetically significant.

The dominant trend of the fold axes and thrusts about the batholith, as shown in Fig. 21.1, is a generalization of the detail shown in Fig. 21.2. The latter map was compiled from the *Tectonic Map of the United States*, with faults of post-Laramide age (as well as known) deleted and with additional fold axes and also some fault detail from the new *Geologic Map of Montana* (1945) added. The conclusion reached by inspection of the detailed map is that the intrusions are markedly discordant locally, but in a broad way the structures of the sedimentary rocks wrap concordantly around the east and north end of the main batholith. This may mean either that the batholith was already there and served as a buttress around which the Laramide structures were wrapped, or that in the process of



intrusion it shouldered aside the adjacent surficial crust and formed the Laramide structures. In the first case the discordant structures would have to be due to later intrusions.

Lewis and Clark "Line"

About 30 miles north of the north end of the Idaho batholith is a zone of large high-angle faults which trends slightly north of west. The chief ones are called the Hope, Osburn, Burnt Cabin, Placer Creek, and St. Joe, and the whole zone referred to as the Lewis and Clark "line" (Wallace *et al.*, 1960). They dominate zones of complexly fractured rock and are the chief localizers of ore in northern Idaho.

The Hope is the most northerly of these great earth fractures. It has a more northwesterly trend than the others, averaging N. 55°–60° W., and dips steeply southwest. It closely parallels the lower course of the Clark Fork of the Columbia River for about 65 miles, then extends through the north arm of Pend Oreille Lake and through a notch across the Selkirk Mountains, giving it a total length of not less than 95 miles. It has an impressive vertical component of movement and stratigraphic throw, but the horizontal component is about 12 miles, the northeast side having been displaced southeast relative to the southwest side. Along the fault zone are many associated fault fractures—low-angle thrust, high-angle reverse, high-angle normal, and two sets of strike-slip faults—all related to the Hope and resulting from the tensional and compressional components of the horizontal shearing stresses which produce the Hope. The faulting, intrusion, and mineralization are closely related events and are regarded as products of the Laramide orogeny.

The Osburn fault of the Coeur d'Alene district is of even greater magnitude than the Hope and has been mapped for 90 miles east-southeast of Coeur d'Alene Lake. Its length is probably much greater, for its course approximately coincides with an old valley extending from Spokane, Washington, to Deer Lodge, Montana, a distance of 300 miles. Its course is N. 70°–80° W. and its dip is steeply south. It also has many associated faults of variable magnitude, some of which are mineralized. Igneous intrusion and mineralization in the Coeur d'Alene district are largely localized along the course of the Osburn fault (Anderson, 1948).

Fig. 21.1. The relation of the Idaho batholith to the Nevadan and Laramide orogenic belts. The Nevadan belt is white and the Laramide belt is dotted and lined. The bold lines in the Laramide belt are axes of prominent folds, thrust faults, and major trends. The Nevadan and Laramide belts overlap; in fact, the geosynclinal division of the Laramide belt was strongly deformed in places in Early and Mid-Cretaceous time.

According to Wallace *et al.* (1960) pronounced strike slip is indicated by the following features:

(a) the offset of large upwarped blocks more or less delineated by areas of outcrop of the Prichard formation, the oldest unit of the Belt series; (b) the offset of major folds and faults, and the dissimilarity of structural features adjacent to one another on opposite sides of the fault; (c) large-scale drag features; (d) offset of the same sense along parallel or subparallel faults; and (e) the position of major mining areas on opposite sides of the Osburn fault and the pattern of ore and gangue-mineral distribution within the areas. A maximum of about 16 miles of right-lateral strike slip is indicated on the segment of the Osburn fault east of the Dobson Pass fault and about 12 miles displacement in the same sense is indicated west of the Dobson Pass fault. The difference in displacement on these two segments is believed to be principally the result of contemporaneous dip slip on the Dobson Pass fault, which has effectively lengthened the block north of the Osburn fault relative to the block south. A few miles east of the area shown in Fig. 21.3, in the vicinity of Superior, Mont., the cumulative lateral movement in the Osburn and the related Boyd Mountain fault, as shown by stratigraphic displacement, appears to be approximately 16 miles, which strongly corroborates the suggested displacement on the Osburn fault in the Coeur d'Alene district.

The age of the Osburn fault is known only within broad limits. It cuts rocks of the Belt Series of Precambrian age and is capped by flows of Columbia River basalt of middle Miocene age. The probably contemporaneous Dobson Pass fault cuts the Gem stocks, which have been dated as about 100 million years old. Other geologic evidence indicates that a lineament in the general position of the Lewis and Clark line may have been in existence since early Precambrian time.

Ages obtained from uraninite from the Sunshine mine indicate that uranium mineralization occurred about 1,250 million years ago. Thus tight folds, such as the Big Creek anticline (Fig. 21.3), that are cut by the uraninite veins, must have been developed before that time. In contrast, the principal ore-bearing veins are younger than the Gem stocks of about 100-million-year age.

The same authors outline the history of development of the structural complex as follows:

During an early stage of deformation (Fig. 21.3A), the principal folds were developed and overturned to the northeast, and reverse faults that strike northwest and dip southwest were formed. A large domelike structure, the Moon Creek–Pine Creek upwarp, was formed west of the reverse faults.

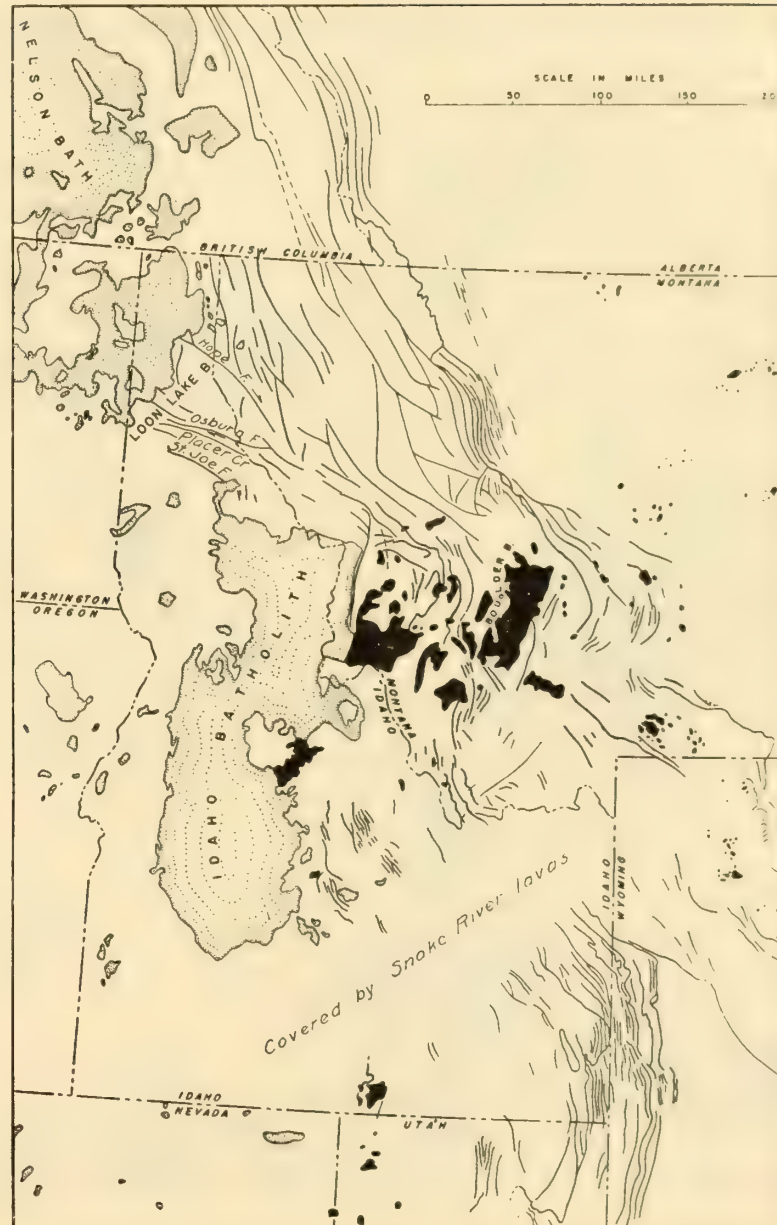


Fig. 21.2. Detail of the belt of Laramide orogeny and the Idaho batholith. Both major fold axes and thrust faults of the Laramide orogeny shown by lines. The main batholith is stippled, and the plutons of known Laramide age are black. Compare with *Tectonic Map of the United States*, 1945.

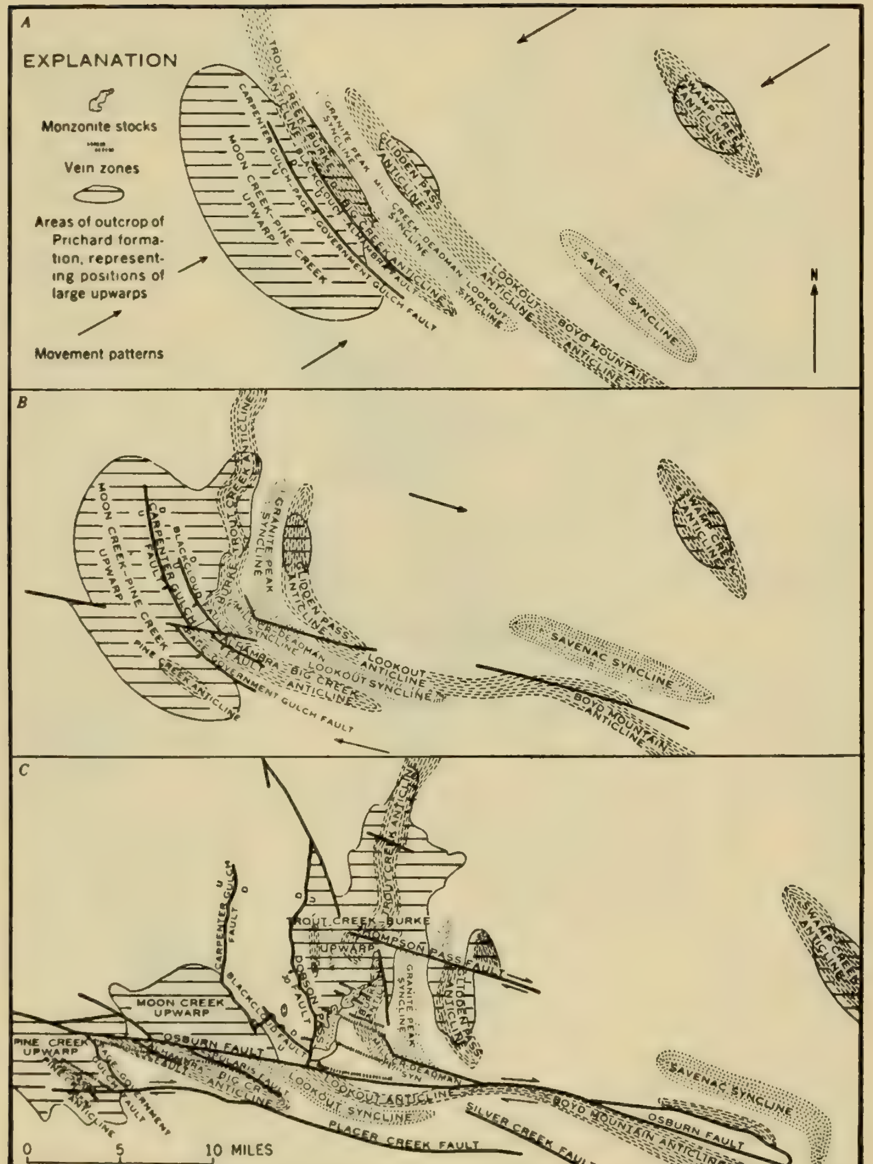


Fig. 21.3. Stages in the development of the Osburn fault zone in the Coeur d'Alene district, Idaho. Reproduced from Wallace et *al.*, 1960.

Accompanying a major reorientation of the stress system, the axes of the folds began to bow (Fig. 21.3B), the southern part of the region moved relatively westward, and incipient strike-slip faults developed. The Mill Creek and Deadman syncline was separated from the Granite Peak syncline and wrapped around the truncated end of the Granite Peak syncline. The northern flank of the Lookout-Boyd Mountain anticline was sliced off by one of the antecedent fractures of the Osburn fault.

Monzonite stocks intruded the structural knot thus produced (Fig. 21.3C), and the principal period of ore deposition followed. Most of the veins are included in spatial groups that define distinct linear belts trending slightly more northwesterly than the Osburn fault system. The concentration in such belts of veins, which are subparallel but differ in size and orientation, suggests that linear feeders for the mineralizing solutions existed at depth, although no through-going structural elements reflect these feeders in the upper crust.

After the principal period of ore deposition, strike-slip movement along the ancestral Osburn zone of weakness became more through-going than previously, and apparently deep-seated stresses were accommodated at this time by displacement on relatively few faults, most of which were in or parallel to this zone. The Osburn fault offset the major folds and early reverse faults, and separated the northern segment of the ore-bearing area from that to the south. The Thompson Pass fault also offset the major folds, and the Placer Creek fault offset the Pine Creek anticline and vein system. The Dobson Pass fault came into existence concurrently with the Osburn fault. The small stocks a few miles west of the Dobson Pass fault may represent cupolas displaced from the main part of the Gem stocks by dip slip on the Dobson Pass fault.

Some of the early-formed tight folds and strike-slip faults were flexed as later rotational stresses were accommodated along newly developed slip planes. Thus, the east end of the Savenac syncline and the adjacent north branch of the Osburn fault were sharply bent and later movement was "short-circuited" along the south segment of the fault. Likewise, the Polaris fault may have accommodated strike-slip deformation after the Placer Creek fault buckled.

Late normal faults, some resulting from the final stages of strike-slip deformation, and others possibly of Quaternary age (Pardee, 1950), have affected the area.

The fault and fold pattern of the map of Fig. 21.2 suggests immediately that the Idaho batholith has moved eastward as a rigid mass, and that the thrusts along its east side are a direct compressional result. But this idea seems incorrect when it is realized that the strike-slip movement on the Osburn fault zone was in the wrong direction.

Setting in Nevadan Tectonic Plan

It is clear that two great arcuate segments of the Nevadan orogenic belt converge in eastern Idaho (refer again to Fig. 17.13), and just a little south of this junction is the Idaho batholith. The same relation to the Laramide orogenic belt has already been pointed out, although the Nevadan segments are convex westward and the Laramide are convex eastward. The Nevadan segments are also curved more and meet at a more acute angle than those of the Laramide. As previously suggested, the junction area of such arcuate segments of a great orogenic belt may be a favorable place for the rise of great batholiths, but it is difficult even to guess why.

The somewhat similar relation of both Nevadan and Laramide belts to the Idaho batholith does not help in restricting or narrowing down the age of the pluton.

Relation to Tertiary Sediments

A fruitful field of research on the age of the Idaho batholith seems to lie in Paleocene and Eocene conglomerates to the east. Certain voluminous conglomerates in northwestern Wyoming are composed of Beltian quartzite boulders and pebbles which are foreign to the formations of the areas in which they occur. Their only source seems to be the Beltian strata that crop out along the eastern edge of the batholith in Idaho and western Montana. See the *Geologic Map of the United States.* Also, Ross (1928) points out that the Idaho batholith was intruded extensively in the Beltian strata, and that a roof of Beltian rocks, fully a mile thick, has been largely removed. In fact, it was removed before the Oligocene and Miocene lavas and sediments accumulated. The connection between the doming of the quartzites, their erosion, and the formation of extensive conglomerate deposits nearby seems obvious; the dating of the intrusions by the conglomerates seems a certain procedure. But the extent and age of the various conglomerates east of the batholith are only fragmentarily known, and some of the conglomerates may be made up of boulders that had already composed a former conglomerate. As far as known, the nearest coarse deposit is the Lima conglomerate in southwestern Montana which is Paleocene in age (Scholten *et al.*, 1955). The oldest of the extensive conglomerates of the Yellowstone–Gros Ventre–Wind River region is late Paleocene in age, and its boulders have been transported a great distance because of the near-perfect rounding of them. This fragment of information suggests very Late Cretaceous or early Paleocene age, again, for the Idaho batholith.

Isotope Age Determinations

The absolute age of the Idaho batholith has recently, and with reasonable assurance, been determined by Larson *et al.* (1954) by lead-alpha activity ratios on the accessory minerals, zircon, monozite, and xenotime. Five analyses yield an average age of 103 m.y. Similar determinations on 7 samples from the Sierra Nevada averaged 100 m.y., and 25 samples from the batholith of southern California gave an age of 105 m.y. Accordingly, it may be concluded that the Idaho batholith is very nearly the same age as the Sierra Nevada. Also, a potassium-argon age determination on the Coast Range batholith near Vancouver by Follinsbee *et al.* (1957) is reported as 105 m.y., again approximately the same. A few years after Larson *et al.* samples were taken by Evernden *et al.* (1957) from 8 individual intrusions in the Sierra Nevada whose age relations had been determined geologically. The samples were run by the potassium-argon method and the ages reported range from 76.9 m.y. for the youngest to 95.3 m.y. for the oldest. These ages are a little under the true absolute age, but not more than a few percent, according to the authors. It may follow that when potassium-argon age determinations are made of the Idaho batholith that they will prove appreciably lower than those of the lead-alpha activity ratio method. Since the Idaho batholith is probably composite, the relation of age determinations by different methods is a bit uncertain, especially since the sequence of intrusions in the Sierra Nevada ranges through 18 m.y.

According to the Holmes B time scale the intrusions dated by the potassium-argon method in the Sierra Nevada range through the Albian (uppermost Lower Cretaceous) and the Cenomanian (lowermost Upper

Cretaceous) (Evernden *et al.*, 1957). Presumably this should be the tentative geologic age assigned to the Idaho batholith. As far as the writer can see there is nothing inconsistent geologically with such a conclusion.

CONCLUSIONS

The Idaho batholith is composite, with some of the smaller parts and satellites of Late Cretaceous and early Tertiary age and some as young as Miocene. The Batholith occurs at the junction area of great arcuate segments of both the Laramide and Nevadan orogenic belts. It is similar in size and composition to the batholiths of the Nevadan orogeny and entirely dissimilar to the plutons of the Laramide belts.

Having intruded the Permian volcanic sequence along its western margin, it lies partly in the Pacific eugeosynclinal province. Its eastern part intrudes miogeosynclinal sediments of the Rocky Mountan type. It is strikingly discordant with the Laramide structures locally, but overall a fairly clear concordance prevails. This and extensive Paleocene conglomerates to the east, derived, presumably, from the roof rock of the batholith, are the best evidence for a Cretaceous age. Isotope age determinations indicate a Mid-Cretaceous date for the main and early components of the great granitic mass. After cooling it formed a buttress against which the Laramide folds and thrusts developed. Still later, younger intrusions cut discordantly through older plutons and the Laramide structures.

22.

CENTRAL ROCKIES

SPATIAL RELATIONS

The system of Laramide mountains referred to here under the heading "Central Rockies" includes the ranges that formed from the geosynclinal sediments of southwestern Montana, eastern Idaho, western Wyoming, central and western Utah, and eastern and southern Nevada. The belt starts at the Idaho batholith and extends southeastward to the Snake River lava plains where it is covered by late Tertiary and Pleistocene lavas and alluvium. See the *Geologic Map of the United States* and Fig 22.1. Emerging from beneath the lavas, it continues southeastward to the Snake River and Hoback ranges of western Wyoming, where it turns southward and extends into northern Utah and to the junction of the east-west-trending Uinta Range. In Utah and Nevada, the Middle and Late Tertiary block faulting has modified somewhat the topographic features resulting from the Laramide orogeny; but it is clear that a belt of complex Laramide thrusting and folding continues on south of the Uinta junction into southwestern Utah and southern Nevada.

The eastern border of the Central Rockies system is sharply defined, whereas the western is indefinite. The eastern margin is made up in part of the Paleozoic strata, in part of the Mesozoic strata and the orogenic deposits of the Cretaceous and the Early Tertiary; but westward only the Paleozoic and some Triassic rocks of the Cordilleran geanticline are involved. Examine the paleotectonic maps of the late Paleozoic and the Mesozoic. Because rocks younger than Paleozoic are almost entirely absent in the western part of the Central Rockies, it is generally impossible to date accurately the phases there or to distinguish the Laramide structures from those of the Cedar Hills, Antler, and Nevadan orogenies. Most probably, the Laramide structures were superposed on the Antler and Nevadan in a medial zone, but details are not known. The map, Fig. 21.1, shows the relation of the orogenic belts to the Laramide as well as possible with existing data.

The Uinta Mountains are a great flat-crested anticlinal uplift and, as far as Paleozoic and Mesozoic strata are concerned, are part of the shelf province. Physiographically, they separate the Colorado Plateau from the great ranges and intermontane basins of Wyoming, and are more closely related to the shelf ranges of Wyoming than to the Colorado Plateau. They are definitely not similar in structure to the Central Rockies, and generally they have thinner formations. Therefore, they are not included in them.

Aside from the Uinta re-entrant in the eastern margin of the central Rockies, the great mountain system is one of approximate arcuate pattern with a radius of curvature of about 450 miles. In it, probably all major overriding thrust sheets have moved continentalward, or toward the convex side of the arc, viz., northeastward and eastward.

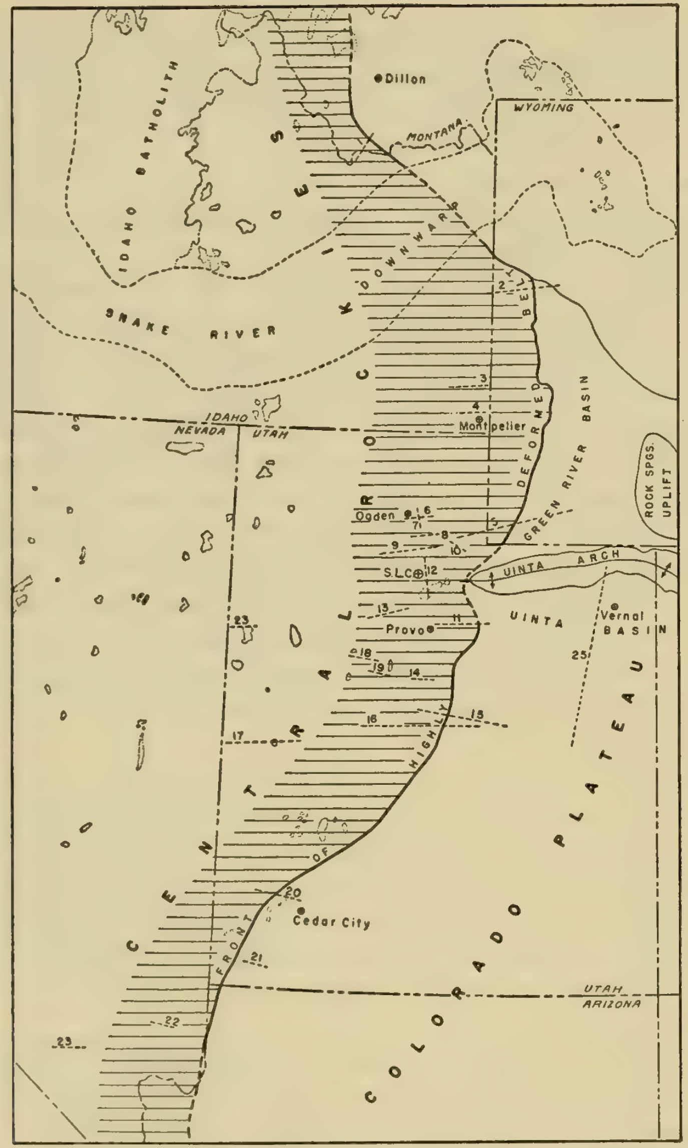


Fig. 22.1. Index map of Central Rockies. Lines of cross sections are indicated by numbers. Intrusive igneous bodies are indicated by dotted lines.

OROGENIC DEPOSITS

A number of coarse conglomerates and thick sequences of sandstone and shale mark the eastern border of the central Rockies, and in connection with thrust faults and unconformities define a succession of orogenic phases. The various formations with which we are mostly concerned from southwestern Montana to southwestern Utah are shown in the correlation chart of Fig. 22.2.

The coarse conglomerates have generally been taken to record the chief phases of mountain building immediately to the west, but thick sequences of sandstone, siltstone, and shale may be equally significant. The conglomerates record settings where a mountain front rose precipitously from a plain, as might have been the case of a vigorously advancing thrust front. But a 10,000-foot section of sandstone, siltstone, and shale of limited time range also records a substantial uplift in the hinterland, possibly less vigorous but sustained, and without an immediately nearby thrust front.

The example of the clastic deposits of Colorado time may be considered (see Fig. 22.3). The lower 3,000 feet of the Indianola group in the Cedar Hills is coarse conglomerate, but eastward and upward it becomes more sandy and shaly. The thick conglomerate has been considered to mark the Cedar Hills orogeny (Chapter 18). An associated thrust sheet rode over part of the conglomerate in the Canyon Range but finally the thrust front was buried by the last of the coarse deposits. Now, going north to the Evanston area an accumulation of more than 8000 feet of sandstone and shale occurs. Conglomerates are insignificant, yet the volume of sediments appears almost as much as in the Indianola area, and the adjacent uplift, therefore, almost as significant.

Where a thrust sheet overlies a coarse conglomerate two orogenic phases might be interpreted; the first to form the conglomerate and the second by the riding of the thrust sheet over the deposit. However, the conglomerate exposed may be simply an early part of the orogenic deposit which was overridden as the thrust sheet advanced, in which case the conglomerate and thrust are manifestations of the same orogeny. Local settings have to be studied individually, and isopach maps such as shown

		SW MONTANA	LIVINGSTON	MT. LEIDY	HOBACK BASIN	EVANSTON	COALVILLE	STRAWBERRY	WASATCH PLATEAU	SW UTAH
PLEISTOCENE		Glacial deps.	?	Glacial deps.	Glacial deps.					
PLIOCENE		Gravels on intermed. surface		Bivouac Teewinot	Camp Davis	Huntsville fangl.			?	Rhyolite flows and pyroclastics
MIOCENE		Medicine Lodge Blacktail Deer Cr.	Bozeman Lake beds	Colter				Bishop cgl.	Traychyte flows	Page Ranch (vols.)
OLIGOCENE		Muddy Cr. Cook Ranch		Wiggins vol.					Gray Gulch vols	Quichapa (vols.) Isom (vols.)
EOCENE		Sage Creek	?	Aycross Wind River Indian Meadows	Pass Pk cgl.	Fowkes tuff Knight	Fowkes tuff Knight	Park City vols. Uinta	Crazy Hollow Green River Colton ?	Needles Range(vols.) Gray Claron
PALEOCENE		Beaverhead cgl.		Pinyon cgl	Hoback	Almy Evanston		Current Cr.	Flagstaff U. North Horn	Red Claron Wasatch
DANIAN	Montana	Undifferentiated (Ruby River Val.)	Livingston	Harebell cgl. Meeteetse		Adaville		Mesaverde	L. North Horn	Kaiparowits
MAESTRICHTIAN			Eagle	Lenticular sequence	?		Echo Can. cgl.		Price River cgl.	Wahweap
SENONIAN				Coaly sequence		Hilliard	Wanship		Blackhawk	Strait Cliffs
TURONIAN	Colorado		Colorado	Bacon Ridge Cody	Frontier	Frontier	Frontier	Mancos	Star Point Indianola gr.(cgl.)	Tropic Dakota Iron Springs
CENOMANIAN				Frontier						
ALBIAN	Lower Cretaceous	Aspen		Mowry	Aspen Bear River	Aspen Bear River	Aspen Kelvin	Mowry Dakota		
APTIAN		Kootenai		Thermopolis	Gannett gr	Gannett gr.	?	Cedar Mountain		
NEOCOMIAN				Cloverly and						
PORTLANDIAN		Morrison	Cloverly ?	Morrison		Morrison	Morrison		Morrison	Morrison

Fig. 22.2. Correlation of Cretaceous and Cenozoic formations along the east front of the Central Rockies.

in Figs. 22.3 to 22.6 compiled in order to understand the situation better.

Although the standard time divisions need not have any bearing on nature's orogenic phases in any particular region they seem to reflect the rhythms or cycles in the Central Rockies. Five main pulses of what are conventionally called compressional orogeny are indicated, namely, Early Cretaceous, Colorado, Montana, Paleocene, and Eocene. The coarse and thick clastic deposits and shifting sites of activity provide the basis for the recognition of the five main phases. After Eocene time volcanism and large-scale normal faulting were widespread and dominant (Fig. 22.7).

SOUTHWESTERN MONTANA

Early Cretaceous Phase

By reference to the paleotectonic maps of Chapter 3, it will be seen that the Paleozoic formations thicken westward into the geosyncline from about Dillon (see Fig. 22.3), and thin to shelf aspects eastward. As an example the Pennsylvanian Quadrant sandstone is nearly 3000 feet thick in the thrust sheet west of Lima, but a few miles to the northeast it is only 400 to 500 feet thick. The shore line of the Triassic and Jurassic formations lay approximately along the Idaho-Montana border west of Lima, but a deep trough failed to develop immediately on the east of the

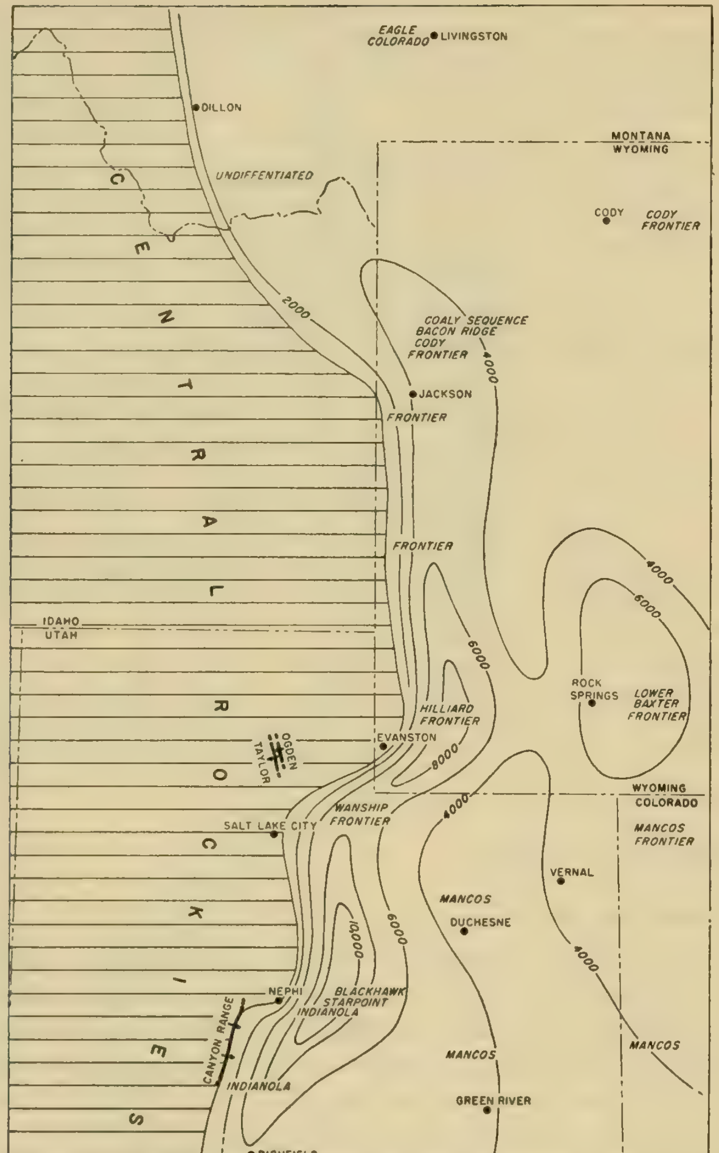


Fig. 22.3. Orogeny and sedimentation in Colorado time in the Central Rockies.

Cordilleran geanticline, as it did in Idaho and Utah, and the sediments of these two periods are only about 2000 feet thick altogether. They thin eastward. Some gentle epeiric movements occurred in the shelf in Jurassic time, as outlined in Chapter 18.

In early Cretaceous time the western geanticline was raised sharply, and the Kootenai conglomerate and arkosic sandstone were washed eastward. It is not thick in most places, but very persistent in western Montana. No structures have been segregated from the structural complex of the west-lying ranges that were formed during the uplift responsible for the Kootenai conglomerate. See paleotectonic map, Plate 15.

A second conglomerate, the Dakota or basal Colorado, is much like the Kootenai; it is perhaps not so uniform in distribution, but it is taken to represent another uplift of the eastern margin of the geanticline.

The radioactivity dates of the Idaho batholith appear to show it a little later than the Dakota conglomerates. At least the first flood of Beltian quartzite boulders so far identified appeared in late Montana time, and these may have arrived at their present destination sometime after the doming of the Beltian strata, consequent to the intrusion of the batholith. The conglomerates in question make up the Harebell formation of the northern Jackson Hole country. These will be considered in a later paragraph.

Montana Phase

Little can be said about events in Montana time in southwestern Montana except that about 5000 feet of sandstone, siltstone, and shale accumulated. These sediments make up an undifferentiated series near Monida, and they undoubtedly attest uplift to the west. See Fig. 22.4.

Paleocene Phase (Mid-Laramide)

In Paleocene time a broad arch of about the size and shape of that of the Big Horn Mountains rose, and extended in a northeast direction (Fig. 22.5). Its southeast flank in part was marked by a thrust fault; its northwest flank was a fairly gentle flexure where observed. The Beaverhead conglomerate seems to be localized around this great arch and to be

made up in large part from Paleozoic limestones and quartzitic sandstones derived from the arch, but in places Beltian boulders are present. These may have come from the west, or from a pre-existing conglomerate not yet found in place. Another uplift west of Yellowstone Park may have appeared at this time, but no conglomerate around it is noted, so the time of the appearance of the uplift and the exposure there of the Precambrian rocks is not yet clear.

The steeply upturned beds and the overriding Precambrian sheet of the southeast flank of the main arch from Lima to Virginia City now stand as the Snowcrest and Green Horn ranges. Part of the northwest flank may be seen in the Blacktail Range southeast of Dillon (Scholten *et al.*, 1955).

The Beaverhead conglomerate is believed to be Paleocene (Lowell and Klepper, 1953). Soon after it was deposited, it was upturned along the Snowcrest Range, and perhaps gently folded in other places.

Early Eocene (?) Phase

We find in southwestern Montana two systems of compressional structures nearly at right angles to each other. The northwesterly trending one is clearly the later (see Fig. 22.6). It is characterized by numerous thrust sheets, some of which override the Beaverhead conglomerate or have carried the conglomerate on their backs in the horizontal movement. See cross section, Fig. 21.8, which runs nearly north-south, just south of Lima. Two folds of the earlier northeasterly trending disturbance are impressed as sharp cross folds in the frontal thrust sheet.

The belt of thrusting of southwestern Montana is undoubtedly a continuation of the one of western Wyoming and eastern Idaho under the Snake River volcanic field, as illustrated in Fig. 22.1. As far as known, all thrusts moved toward the northeast. One or two brought the Precambrian crystalline rocks to exposure, but now are dismembered by erosion into klippen and fensters. They involved the Paleozoic rocks of geosynclinal character, and along the eastern front of the thrust belt the Mesozoic rocks occur and are deformed.

The belt of thrusting on the northeast, between Virginia City and Bozeman, involved the thin-shelf sediments, and in the uplifting that accompanied each thrust sheet, much of the Paleozoic and Mesozoic

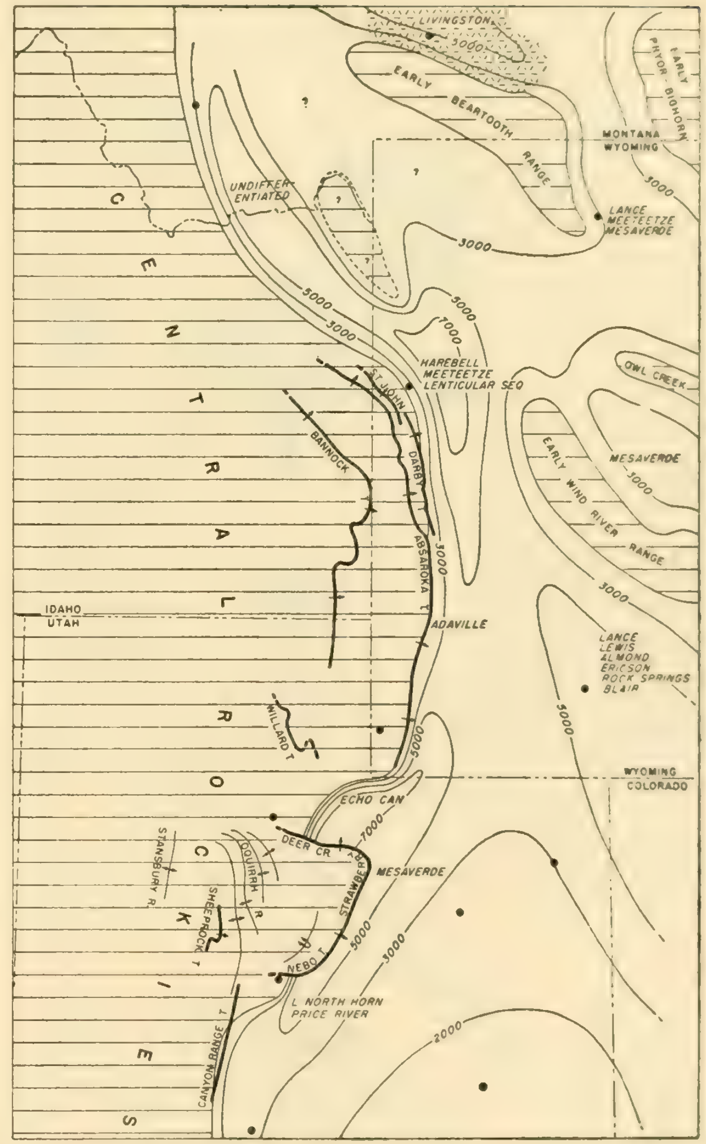


Fig. 22.4. Orogeny and sedimentation in Montana time in the Central Rockies. Absaroka is thrust over Adaville.

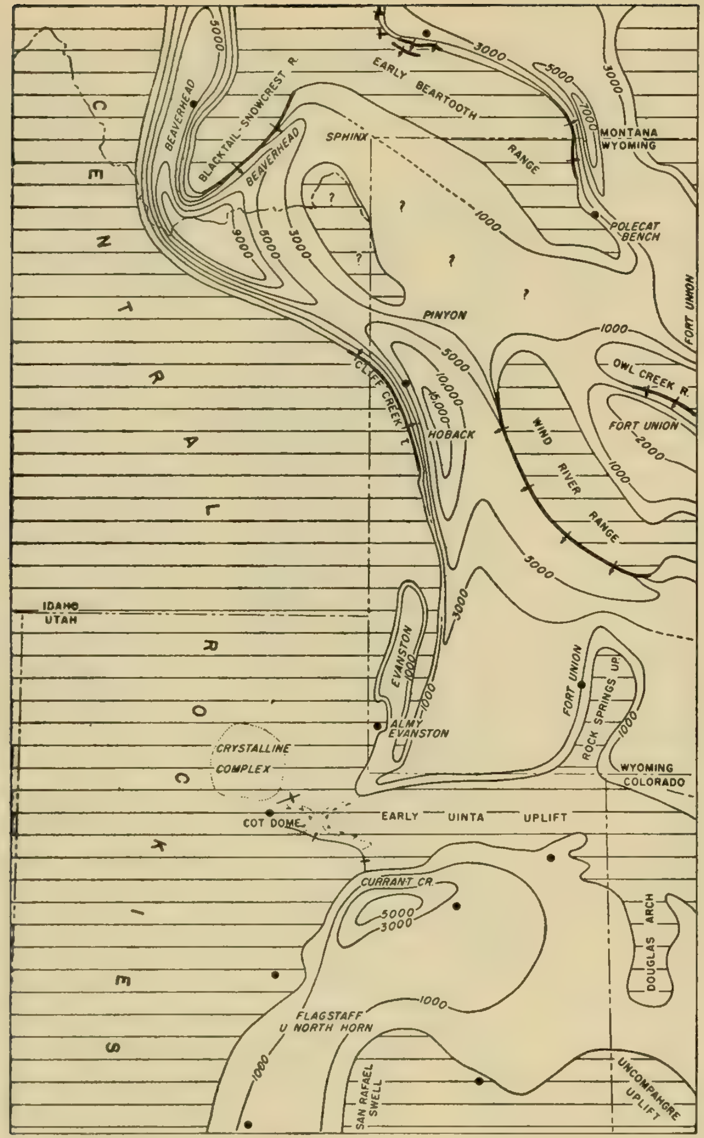


Fig. 22.5. Orogeny and sedimentation in Paleocene time in Central Rockies. Lower part of Evanston formation is latest Cretaceous.

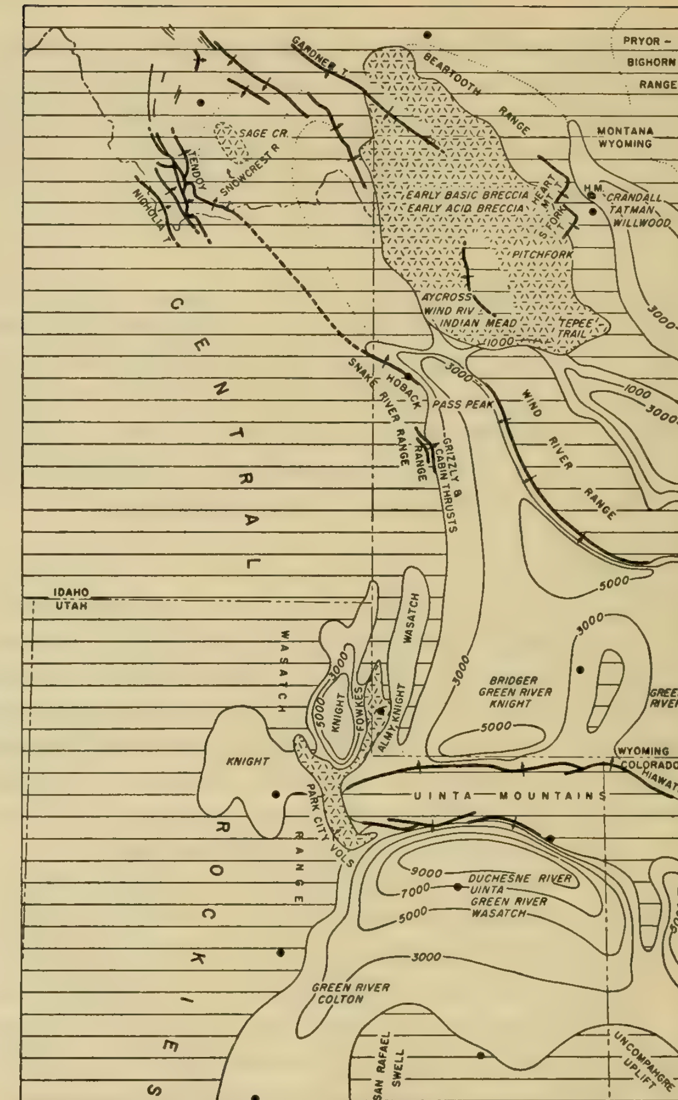


Fig. 22.6. Orogeny and sedimentation in Eocene time in the Central Rockies. Heart Mountain and South Fork thrusts shown at place where they originated. H.M. is Heart Mountain, a glide block of the Heart Mountain thrust.

veneer was removed, and the crystalline rocks were exposed. The thrust sheets dip fairly steeply in this belt both northeasterly and southwesterly.

Numerous folds in the Paleozoic and Mesozoic strata developed at the same time as the thrusting. The Beaverhead conglomerate was much eroded after the thrusting and folding, and a singular remnant, the Sphinx conglomerate, now holds up the highest peak in the Madison Range, Sphinx Mountain, northwest of the northwest corner of Wyoming.

Several porphyry stocks were intruded immediately after the thrusting along the Idaho-Montana border in the Nicholai and Cabin thrust sheets (Fig. 22.6), and it is probable that a good deal of the intrusive and mineralizing activity in the Melrose, Butte, and Philipsburg areas, immediately to the north, occurred at this time.

Late Eocene to Early Miocene Phase

Following the main thrusting in southwestern Montana, a long episode of erosion, with possibly some additional crustal movements, changed the topography to an almost unrecognizable extent. The arches and thrust sheets that had brought Precambrian rock to exposure were irregularly reduced, and perhaps broadly downfolded in places. Instead of concentrating their attack on the sedimentary rocks, the erosional processes cut great intermontane valleys through the Precambrian crystalline rocks as well, with only local structural control.

Then, in late Eocene time, volcanism broke out in nearby regions, and focused in Yellowstone Park and the Absaroka Range (Fig. 22.8). Volcanism of superior magnitude also broke out in the Coast Range region of Oregon and Washington at this time. It resulted in the damming of drainage ways and in abundant ash and dust falls. The regimen of erosion changed to one of alluviation in the great intermontane valleys, and the heavy deposition of the Sage Creek formation (late Eocene) resulted in southwestern Montana. Other formations of equivalent age were laid down in the basins elsewhere over a wide region.

Local deformation and erosion in early Oligocene time are noted by an unconformity between the Sage Creek beds and those that overlie it. Volcanism continued nearby, and the deposition of the Cook Ranch beds in middle Oligocene time on the Sage Creek beds resulted.

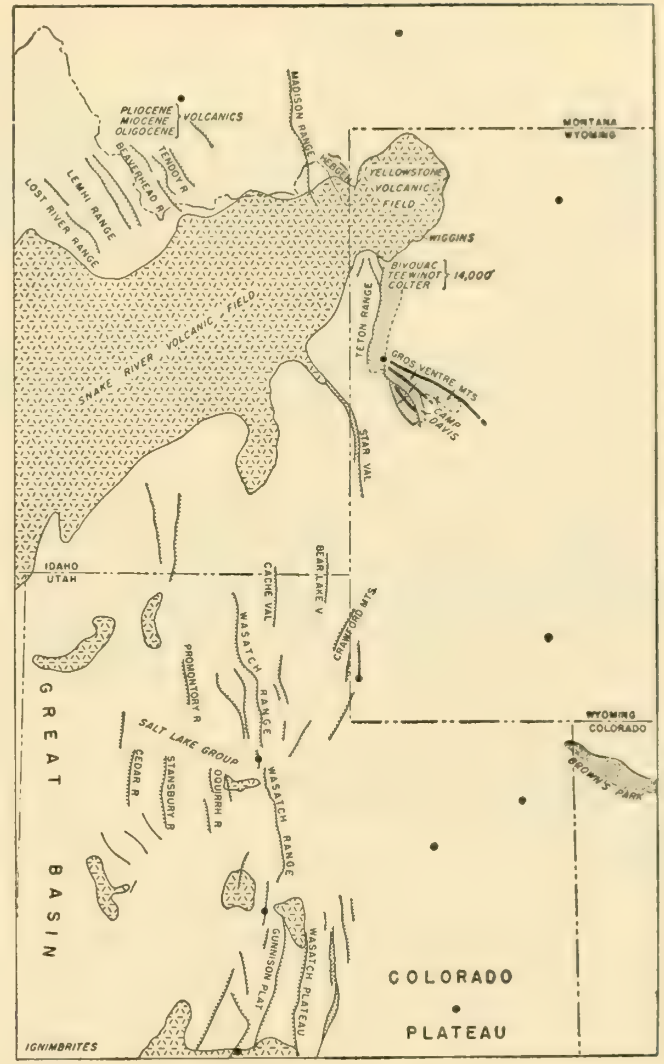


Fig. 22.7. Orogeny, sedimentation, and volcanism in late Cenozoic time in the Central Rockies. The Great Basin is brought into existence by block faulting and becomes a region of considerable sedimentation (the Salt Lake group) on the downfaulted blocks. The east end of the Uinta Mountains sank along the axial area and the Browns Park formation was deposited in the depression. Volcanism starting in late Eocene and running through the Cenozoic was widespread. Much of the Tertiary volcanic rocks in the Geat Basin are buried by later alluvium.

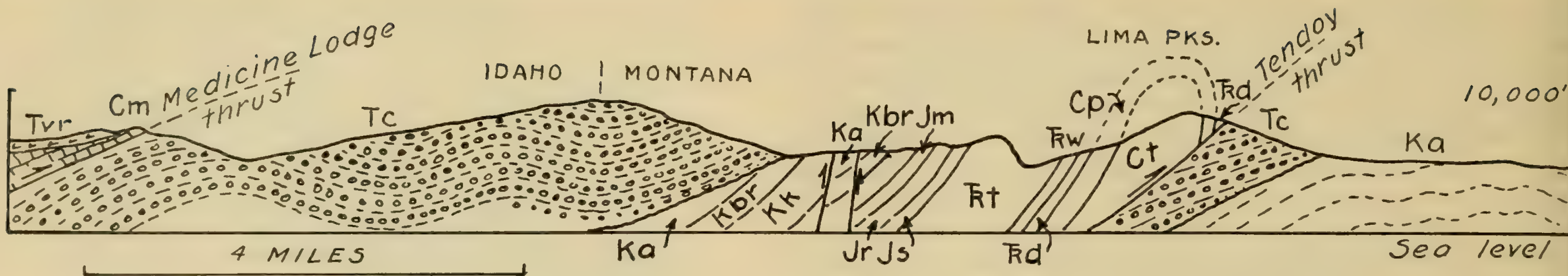


Fig. 22.8. Cross section of thrusts in southwestern Montana and adjacent Idaho, after Drexler, McUsic, and Kildal, Master's theses, University of Michigan. Cm, Madison formation; Ct, Tensleep sandstone; Cp, Phosphoria formation; Rd, Dinwoody formation, Ꞧw, Woodside formation; Ꞧt, Thaynes formation; Js, Sawtooth formation; Jr, Rierdon formation; Jm, Morrison formation; Kk, Kootenay formation; Kbr, Bear River formation; Ka, Aspen formation; Tc, Paleocene (?) conglomerate; Tvr, Rhyolite flows. Section 1, Index map, Fig. 22.1.

Miocene–Pliocene Phase

A fairly extensive episode of erosion followed the deposition of the Cook Ranch beds, and in the Blacktail Range southeast of Dillon, tilting and the overlap of younger beds seem to indicate the inception of block faulting. This would have occurred in latest Oligocene or earliest Miocene time. Then volcanism broke out anew at the north end of Blacktail Range and extensively in the Snake River Valley, Yellowstone Park, and the Columbia Plateau. Deposition of lower Miocene Blacktail Deer Creek beds and associated basalts, tuffs, and agglomerates resulted in the Upper Sage Creek area, along the northwest flank of the Snowcrest Range, and in the Ruby Reservoir basin.

Then followed erosion to an extensive surface of moderate relief. In places the pre-Sage Creek surface may have been reëxhumed and become coextensive with this new post-Blacktail surface, which is present now in summit areas of the Blacktail Range. There, lower Miocene basalts and tuffaceous beds are beveled.

An episode of block faulting is clearly recorded in the Ruby Reservoir basin following the deposition of the Blacktail beds, and then in the downfaulted basin, the upper Miocene and lower Pliocene Madison Valley beds accumulated.

Pliocene and Quaternary Faulting and Erosion

Regional uplift, in places possibly accompanied by more block faulting, and the erosion of extensive pediments followed. The pediments on the northwest side of Snowcrest Range are the most extensively and perfectly developed. The pediments on basin beds of the back valleys in Beaverhead Range (graben valleys) are of this age. In valleys like Beaverhead River, Blacktail Creek, and Sweetwater, downfaulting was so extensive that alluvial aprons were deposited along the base of the fault scarps.

A third episode of block faulting resulted in alluviation in places, and in others of gentle uplift and dissection of the pediments. Two episodes of glaciation in the Beaverhead Range are recorded, one probably occurring before dissection of the pediments, and one afterward.

Block faulting at the front of the Tendoy and Madison ranges has continued in modern times.

SOUTHEASTERN IDAHO AND WESTERN WYOMING

Latest Jurassic and Early Cretaceous Phase

Like the southwestern Montana Rockies, those of southeastern Idaho and western Wyoming contain Paleozoic formations of geosynclinal thicknesses on the west, of shelf thicknesses on the east, and of marginal ge-

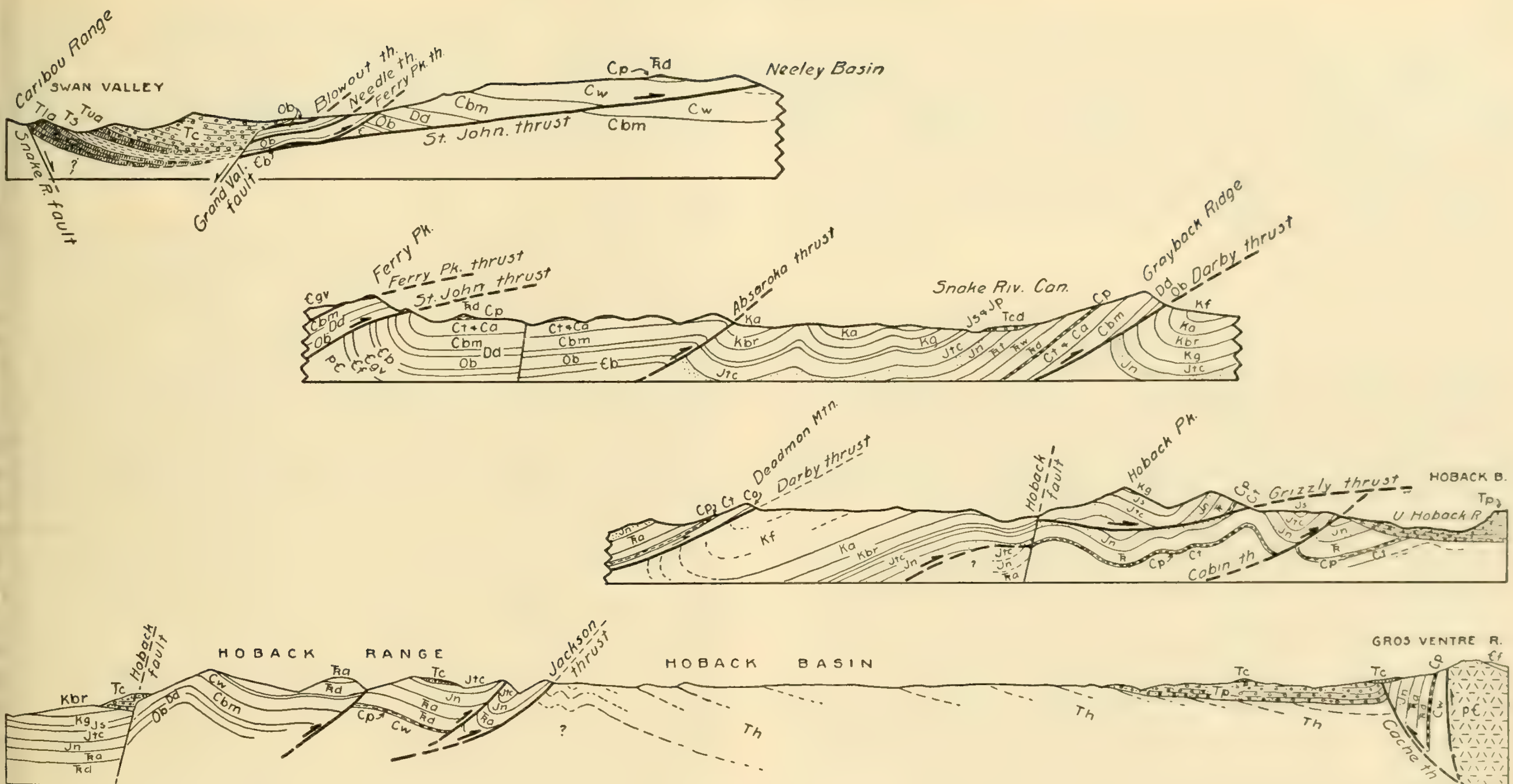


Fig. 22.9. Cross sections of the northern central Rockies from the Caribou Range in Idaho eastward to the Hoback basin (north end of the Green River basin) in Wyoming. See section 2, Fig. 22.1. The sections are not continuous but each is staggered southward somewhat from west to east. (Upper section adapted from R. Enyert's thesis; middle section adapted from K. Keenmon's thesis; lower section adapted from Jack St. John's and Alex Ross' thesis, all of the University of Michigan.) The Ferry Peak, Absaroka and Darby thrusts are Paleocene in age; the Cabin thrust is late lower Eocene (post-Hoback fm.); and the Grizzly thrust is late middle Eocene (post-Pass Peak congl.). Also a cross section from Hoback Range to the Gros Ventre Range across the Hoback basin, which is the north end of the Green River basin. Ꞓf, Flathead quartzite; Ꞓgv, Gros Ventre formation; Ꞓb, Boysen formation; Ob, Bighorn dolomite; Dd, Darby formation; Cbm Brazer and Madison limestones; Ca, Amsden formation; Ct, Tensleep sandstone; Cw, Wells (Amsden and Tensleep); Cp, Phosphoria fm; Ꞧd, Dinwoody and Woodside; Ꞧa, Ankareh; Jn, Nugget sandstone; Jtc, Gypsum Spring and Twin Creek; Js, Preuss and Stump; Kg, Gannett group; Kbr, Bear River; Ka, Aspen; Th, Hoback fm.; Tp, Pass Peak congl.; Tc and Tcd, Camp Davis fm.; Tla, lower andesite, Tua, upper andesite, Ts, silt of Camp Davis; Kf, Frontier fm.

anticlinal deposts of Mesozoic age in their central and eastern parts. Refer again to the paleotectonic maps of Chapter 3.

The Ephraim conglomerate marks the first vigorous uplift of the geanticline to the west, and the age of the conglomerate, according to Mansfield (1927), is Early Cretaceous, but according to W. L. Stokes (personal communication) may be latest Jurassic. Somewhat later, but still in early Cretaceous time, the Bechler conglomerate was washed eastward from the westward-lying geanticline.

Colorado Phase

The orogenic deposits of the Colorado phase (Fig. 22.3 north and northeast of Jackson are the Frontier formation, Cody shale, Bacon Ridge sandstone, and the Coaly Sequence (Love, 1956a,b). They make up a series of clastic deposits about 5000 feet thick. East of Evanston the Frontier formation and Hilliard shale are about 9000 feet thick. These deposits undoubtedly attest the rise of adjacent land on the west, but for most of the length of the deformed belt it is impossible to identify any structures there that were formed at this time. The Taylor and Ogden thrusts predate the Willard thrusting, which is probably Montana in age, so they may be structures formed as the west-lying land was elevated.

Montana Phase (Early Laramide)

The deformed belt of western Wyoming and southeastern Idaho is noted for a number of thrust faults, the main ones of which are shown on Fig. 22.4. They have all moved eastward, or at the north end of the belt northeastward, and in places a number of sheets are stacked on each other in imbricate fashion. These probably formed during late Montana or early Paleocene time.

The Bannock thrust was first detailed by Mansfield (1927) as shown in Fig. 22.10. A sheet of wide proportions was postulated to have moved eastward over 40 miles and to have been folded and eroded such that a large window occurs in it. Later work by geologists of Standard Oil Company of California and the U.S. Geological Survey indicates that several imbricate thrust sheets are involved and that the interpretation of one single sheet is not correct.

The Absaroka thrust has been traced the entire length of the belt and is an integral part of the frontal structure of the central and southern parts. To the north it runs back of the Darby thrust, presumably of the same age. Also on the north end a complex of thrust sheets, one particularly of considerable extent, the St. John, overrides the Absaroka in the Snake River Range. It may belong to the Paleocene or Eocene phase of deformation.

The youngest strata involved in the Bannock thrusting are Lower Cretaceous Gannett. The Frontier formation of Colorado age is deformed within the Absaroka and Darby sheets. The Absaroka overrides the Adaville beds near Kemmerer.

The foredeep beds deposited during the Montana epoch attained a thickness of over 7000 feet in the Jackson area, and their deposition climaxed in the Harebell conglomerate of Beltian boulders, cobbles and pebbles (Love, 1956a). At the extreme southern end of the belt the Echo Canyon conglomerate and related deposits accumulated at about the same time (Williams and Madsen, 1959). The manner and route of long-distance transit of the Beltian cobbles of the Harebell conglomerate from closest Beltian outcrops 200 miles to the northwest are a mystery.

Since the Adaville is overridden by the Absaroka thrust, the deformation, at least here along the front of the belt of deformation, carried on into late Montana time and possibly into early Paleocene. The Paleocene Hoback formation was deposited in a foredeep (see map, Fig. 22.5), and it is possible that the foredeep occurred in response to the thrusting, and that the thrusting is therefore related to the Hoback formation rather than to the late Montana sediments. The thrusting in the southern end of the belt is pre-Knight, and Veatch (1907) had presumed it to be post-Almy, but a recent revision of the stratigraphy and mapping in the Fossil basin (Tracy and Oriel, 1959) shows the thrusting there to be pre-Evanston. The lower part of the Evanston is latest Cretaceous, and hence the thrusting is Late Cretaceous. Deformation in and around Fossil basin continued through the Paleocene, however, as indicated by the conglomerates and unconformities in the Evanston and Almy.

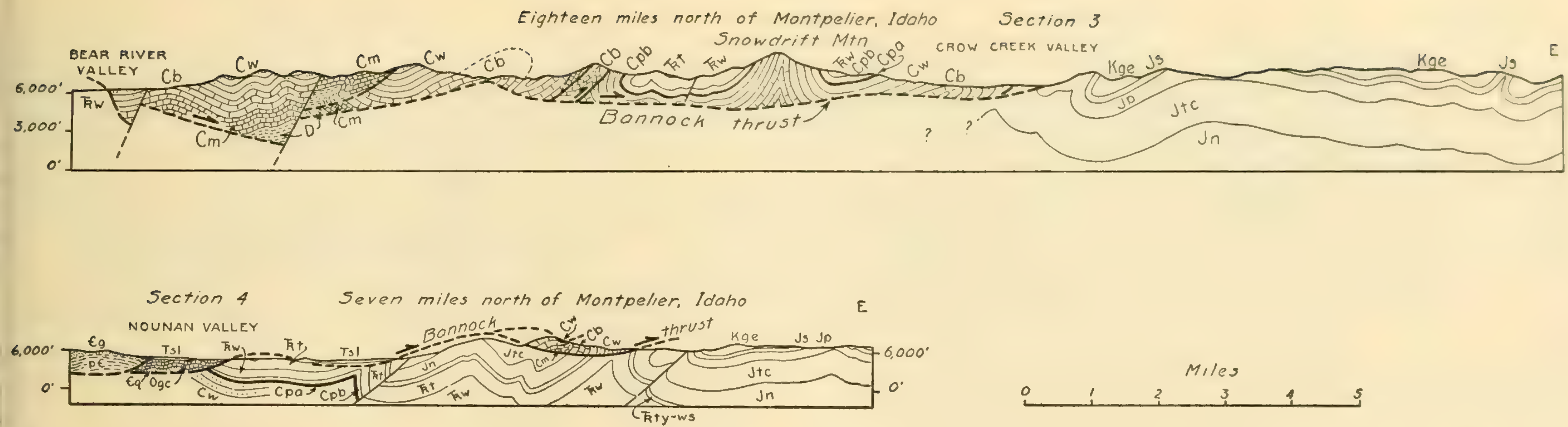


Fig. 22.10. Cross sections of the Central Rockies in southeastern Idaho, after Mansfield, 1927. Ꞓq, Brigham quartzite; Ogc, Garden City ls.; D, Devonian Three Forks or Jefferson ls.; Cm, Madison ls.; Cb, Brazer ls.; Cw, Wells quartzite; Cpa and Cpb, Phosphoria fm.; TRw, Woodside sh.; TRt, Thaynes Group; TRty, Timothy sandstone; TRh, Higham grit; TRd, Deadman ls.; Jn, Nugget ss.; Jtc, Twin Creek fm.; Jkb, Beckwith fm.; Jp, Preuss ss.; Js, Stump ss.; Kge, Ephram conglomerate of the Gannett group.

The Bannock and Willard thrusts are presumed to have formed in Montana time, the same as the Absaroka, but they might be older.

Figure 22.11 shows thrusting during the deposition of the Echo Canyon conglomerate, but this is an inferred structure.

Paleocene Phase (Mid-Laramide)

Figure 22.5 illustrates deposits and uplifts along the east front of the Central Rockies in Paleocene time. The major sediment accumulation was the continental Hoback formation made up of about 15,000 feet of sandstone, siltstone, and shale. A few thin limestone and conglomerate beds are also present (Dorr, 1958).

Sedimentation was very rapid, probably beginning and accelerating in Torrejonian, culminating during late Torrejonian, then decelerating during Tiffanian, Clarkforkian, and Graybullian times prior to a late phase of orogeny. Sediment was derived locally from western, mid-Laramide highlands which began to rise in the early Torrejonian; the uplift culminated between the middle and end of the Torrejonian. Orogenic phases were relatively brief but intense. The area of deposition was much lower, forested, temperate, humid, locally swampy with some lakes, and largely inhabited by a forest-dwelling mammalian fauna (Door, 1958).

The Cliff Creek thrust sheet (Jackson thrust of Fig. 22.9) overrides the Hoback formation. It is overlapped by the Eocene Pass Peak formation, so probably is a last phase of the deformation during the Paleocene which resulted in the deposition of the Hoback beds.

The Uinta uplift appeared first in Paleocene time. The Currant Creek conglomerate, which had previously been related to the Montana Price River of central Utah, is now regarded as Paleocene by Bissell (1959). It rests unconformably on older strata, and postdates the Deer Creek–Strawberry thrusts.

The linear uplift and basin (Fossil basin) east and north of Evanston are developments during latest Montana-Paleocene time.

Eocene Phase (Late Laramide)

At the north end of the deformed belt of western Wyoming the Pass Peak conglomerate was deposited on the Hoback formation and older

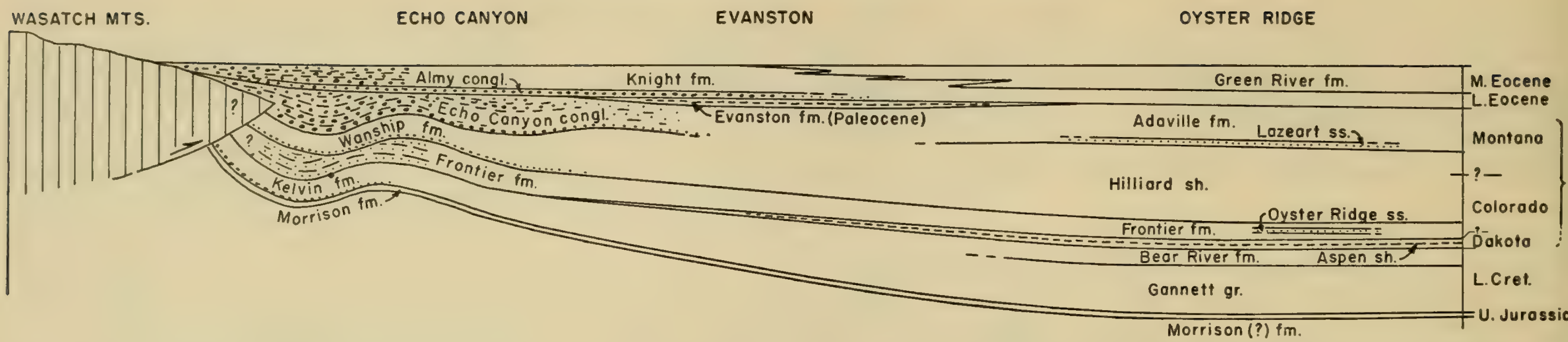


Fig. 22.11. Idealized cross section showing relations of formations from Wasatch Mountains to Oyster Ridge, Wyo., restored to the close of Green River deposition. Folding of Knight near Wasatch Mountains followed, then erosion, then deposition of Norwood tuff (early Oligocene), the Basin and Range type faulting. Section 5, Fig. 22.1.

beds, and apparently immediately overridden by rising thrust sheets both on the east and west (Fig. 22.6). The Wind River Range rose along a high-angle thrust which cuts the conglomerate, and new thrusts broke out in the Hoback Range on the west (Grizzly and Cabin thrusts, Fig. 22.9). The Cliff Creek thrust sheet is overridden by these later slices.

The Knight formation was spread widely over erosion-beveled strata in the Evanston–Salt Lake City area, and then was itself folded in large open folds with amplitude of several thousand feet and fold widths of 5 to 15 miles. The frontal Wasatch Range north of Salt Lake City first came into existence at this time. The folding was accompanied by longitudinal normal faulting (Eardley, 1944), and this may have marked the inception of Basin and Range faulting.

In very late Eocene time the Park City volcanic field was formed, and the related Fowkes tuff accumulated in erosional valleys in the Knight conglomerate and older formations.

The Uinta Mountains had their chief growth in late Eocene time (Bridger, Uinta, and Duchesne River time).

Late Cenozoic Phases

The chief orogenic activity in late Cenozoic time was block or rift faulting. A belt of trenches, horsts, and tilted blocks formed from northwestern Arizona through western Wyoming and southeastern Idaho to British Columbia, and ranges and valleys came into existence such as shown on Fig. 22.7.

At the junction of the fold belt with the Wind River–Gros Ventre uplift overthrusting occurred in early Pliocene time (Love, 1956b). This late thrusting is unique in the Rocky Mountains and most probably does not represent a part of an extensive compressional belt. We have to deal with the deposition of the Camp Davis conglomerate, overthrusting on the conglomerate, and normal block faulting all in a very short time. The thrusts are also not traceable for any appreciable distance. These observations lead the writer to the conclusion that the thrusting is a gravity slide phenomenon associated with uplift. The events, structures, and deposits would be interpreted as follows. Normal faulting of vigorous nature started, and on the down-thrown block the conglomerate accumulated to a thickness of 2000 to 3000 feet. Then gravity gliding of large masses from the upthrown blocks occurred down over the conglomerate in places. Deposition of conglomerate continued around the slide masses, and with continued normal faulting the slide masses across the fault were cut and offset, and the uplifted parts removed by erosion. As the fault pattern is studied it seems to fit best, if not require, this interpretation.

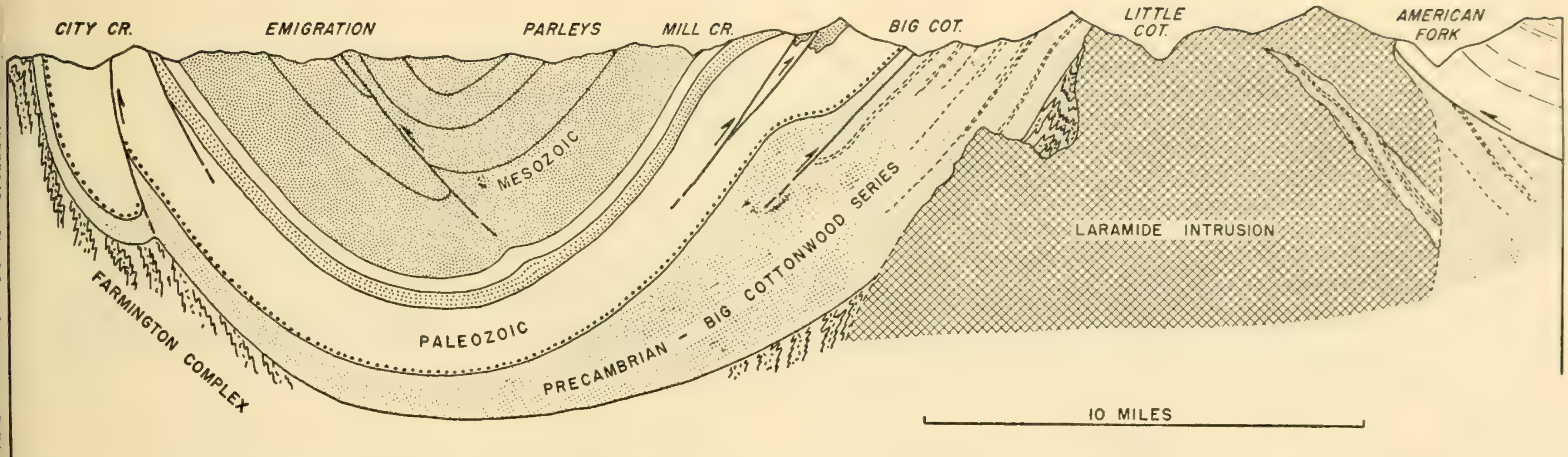


Fig. 22.12. Cross section lengthwise of the Wasatch Mountains east of Salt Lake City. After Granger, Sharp, and Crittenden, unpublished map. Section 12, Fig. 22.1.

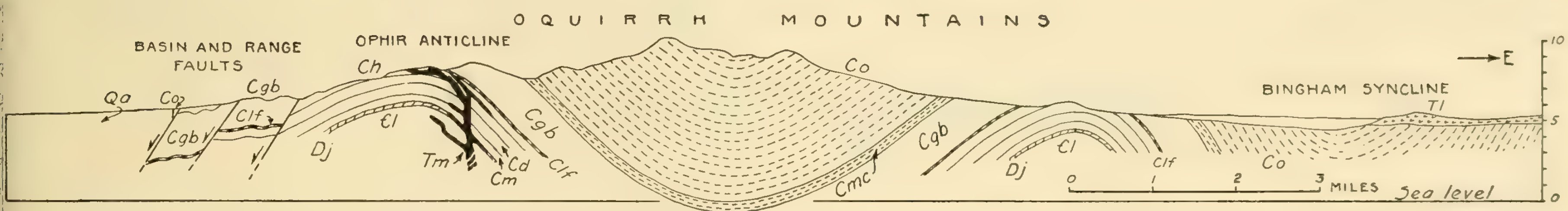


Fig. 22.13. Cross section of the Oquirrh Mountains. After Gilluly, 1932. Section 13, Fig. 22.1.

WASATCH AREA OF UTAH

Colorado Phase (Cedar Hills Orogeny)

The foredeep basin east of Evanston continued to the southwest, and in the Wasatch area east of Salt Lake City the Frontier and Wanship formations were deposited in it, making up a sequence of sandstones, shales, and coal beds about 7000 feet thick. See Fig. 22.3. A conglomerate about 50 feet thick forms the lower part of the Wanship which rests unconformably on the Frontier and older beds.

Although the degree of discordance is very slight and difficult of recognition in the Coalville area, the uncomformity is very pronounced locally and attains 90° of discordance at the head of Dry Creek about 2 miles east of Rockport Reservoir. The angular unconformity is at the base of the conglomerate that is near the middle of the sequence in the Coalville area which has heretofore been regarded as Frontier. At the Dry Canyon locality the conglomerate con-

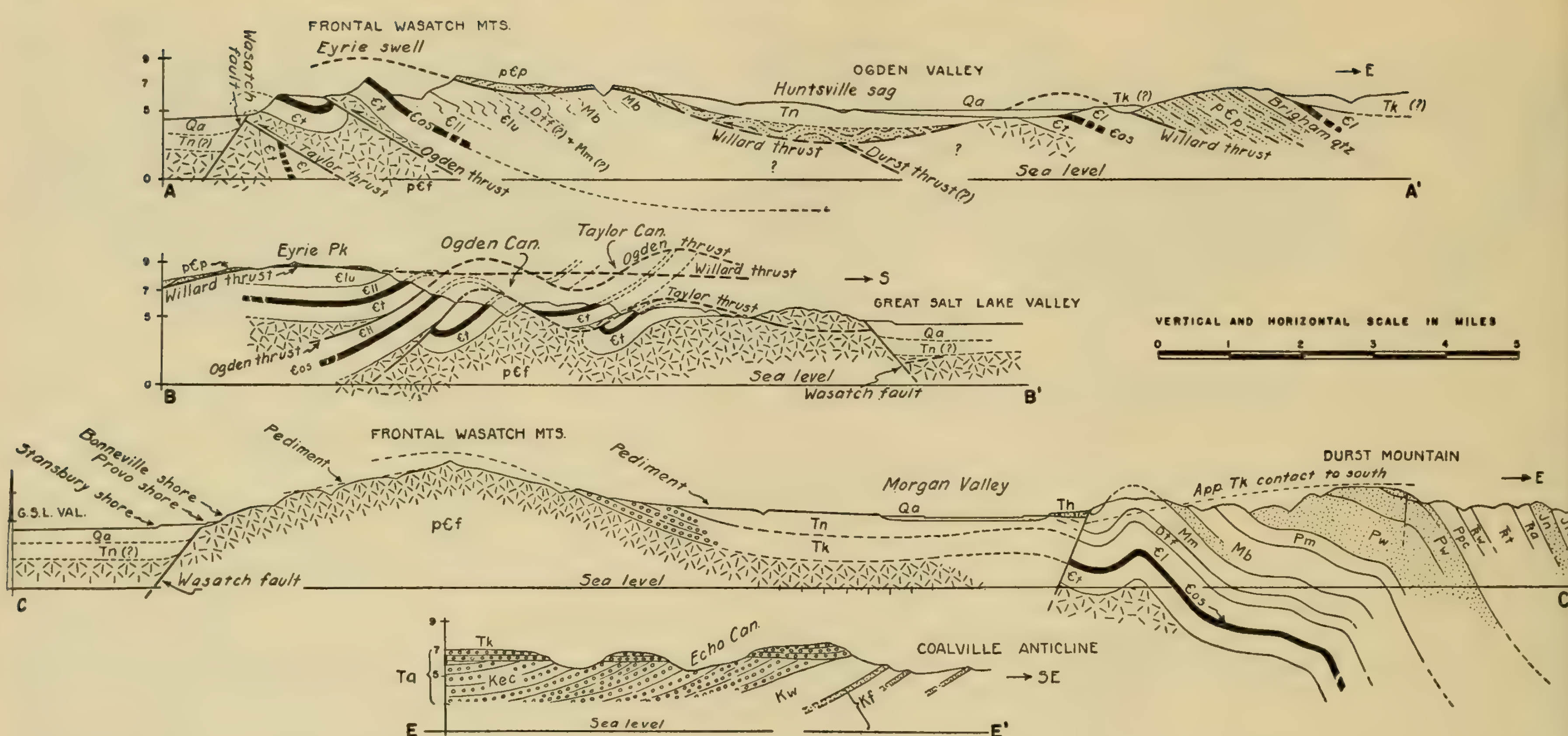


Fig. 22.14. Cross sections of the north-central Wasatch Mountains, after Eardley, 1944. A–A′, an east-west section just south of Ogden Canyon extending from the Great Salt Lake to Ogden Valley and beyond (section 6). B—B′, a north-south section in the range from the north to the south side of Ogden Canyon. The section terminates in the Great Salt Lake Valley because the mountain front veers to the east of this place (section 7). This shows the crossfolding of the Taylor and Ogden thrust sheets. C—C′, an east-west section from the Great Salt Lake Valley to Durst Mountain about midway between Salt Lake City and Ogden (section 9). Formation; p€f, Farmington Canyon complex; p€p, Proterozoic strata; €t, Tintic quartzite; €os, Ophir shale; €l, €lu, €ll, Cambrian limestone, upper and lower divisions; Dtf, Three Forks (?) formation; Mn, Madison limestone; Mb, Brazer formation; Pm, Morgan sandstone; Pw, Weber quartzite; Ŧw, Woodside shale; Ŧt, Thaynes limestone; Ŧa, Ankareh formation; Jn, Nugget sandstone; Je, Entrada sandstone; Jte, Twin Creek formation; Kk, Kelvin formation; Ka, Aspen fm.; Kf, Frontier formation; Kw, Wanship fm.; Tk, Knight formation; Tn, Norwood tuff (latest Oligocene, now same as Fowkes).

tains boulders at least as old as the Gardison (Madison) Formation. Between Dry Creek and Crandall Canyons the conglomerate rests variously upon late Jurassic, Early Cretaceous and older late Cretaceous strata and lies undisturbed across two post-Carlile faults. Thus, a marked though perhaps localized tectonic disturbance is indicated (Williams and Madsen, 1959).

The faulting and folding of post-Frontier and pre-Wanship time are part of the Cedar Hills orogeny which centered farther south. At first they were thought to mark an early uplift of the northwest end of the Uinta Mountains, but when one isopachs the formations of Colorado age

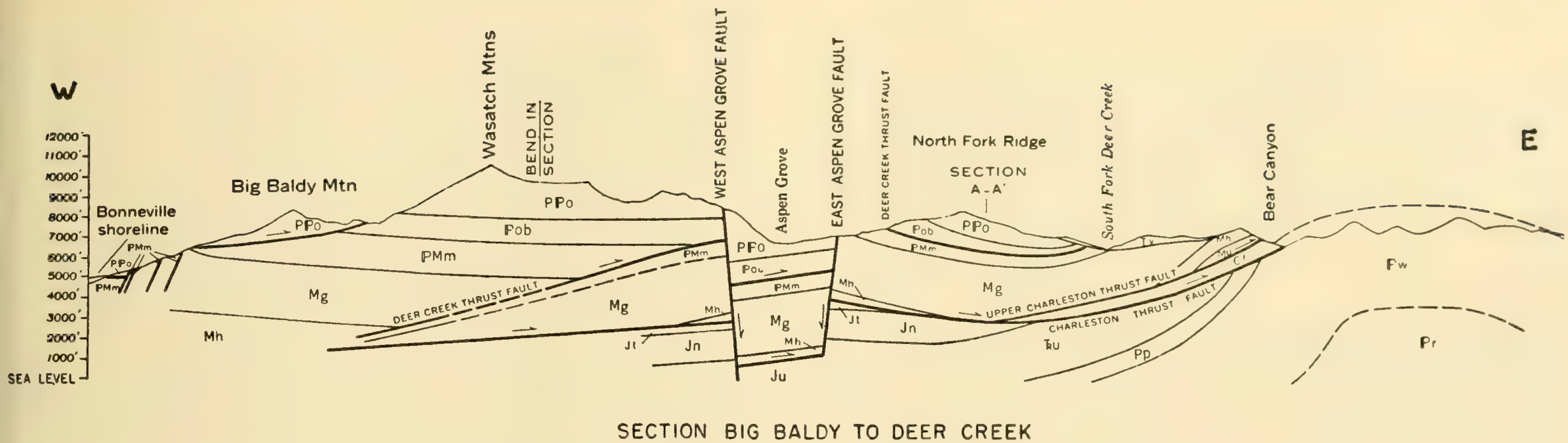


Fig. 22.15. Cross section of Wasatch Mountains east of Provo. Reproduced from Baker, 1959.

and considers their lithology, this surmise seems doubtful—no uplift in the site of the Uinta Mountains appears to have existed at this time.

In the Ogden segment of the Wasatch Mountains the Taylor and Ogden thrusts were formed as indicated in section A–A′ of Fig. 2.14, sometime preceding the Willard thrust. The Willard thrust is considered Montana in age, so therefore the Taylor and Ogden thrusts are probably Colorado. They were then cross folded with axes trending east-west, as shown in section B–B′. The cross section may also be interpreted to mean that the beds were folded before the thrusting. Since the cross-folding involves a different framework of stresses it seems probable that the two were formed some time apart. However, for the present they will both be considered to have developed during the Colorado epoch.

Montana Phase (Early Laramide)

The deposition of the Echo Canyon conglomerate (Williams and Madsen, 1959) and a related sequence east of Henefer (Eardley, 1944) over 8000 feet thick marks a major phase of orogeny immediately to the west. It began in latest Colorado time and ran its course well into the Montana epoch. The Willard thrust seems to have formed at this time as well as the major Charleston–Deer Creek–Strawberry–Nebo line of thrusts. Figure 22.15 by Baker (1959) shows the extensive, flat-bottomed thrust sheets of the Provo section of the Wasatch Mountains, and Fig. 22.16, the Nebo thrust at the south end of the Wasatch Mountains. The interpretation rendered in Fig. 22.4 suggests a great gravity slide for the thrust salient. It may be noted that the sheet is made up largely of the thick Oquirrh formation of Pennsylvanian age, described in Chapter 6.

The broad folds of the Oquirrh and Great Basin ranges nearby, as well as the Sheeprock thrust (Cohenour, 1957) possibly formed in Montana time. See Fig. 22.13.

Paleocene Phase (Mid-Laramide)

The Currant Creek conglomerate (Fig. 22.5) rests unconformably across the beveled edges of older formations at the southwest end of the Uinta Mountains and although not definitely dated paleontologically is called Paleocene by Bissell (1959). It marks the first rise of the Uinta

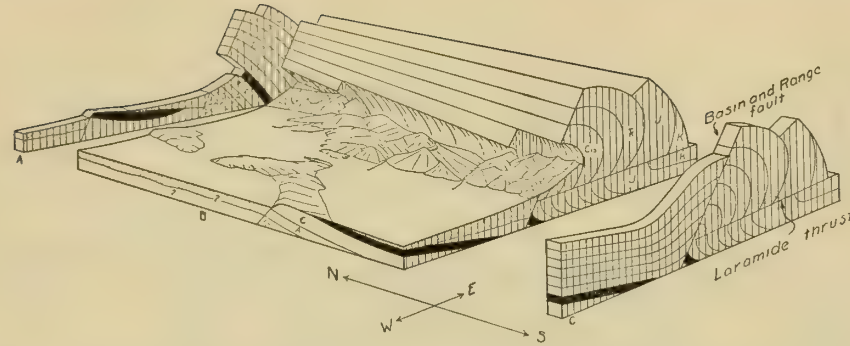


Fig. 22.16. Diagram of the southern Wasatch Mountains showing the Nebo thrust and the later normal block faulting. After Eardley, 1933. Section 14, Fig. 22.1.

uplift. It is not clear what relation the Evanston formation on the north bears to the early Uinta uplift. The Cottonwood dome and east-west folding north of it also probably formed at this time. See Fig. 22.12. In north-south cross section the Cottonwood intrusion appears to have had some effect on the doming, but it is quite discordant eastward and may be later. It is discussed further in Chapter 36.

Eocene Phase (Late Laramide)

The Knight conglomerate of early Eocene age was spread widely over deformed and eroded strata of the east front of the central Rockies and marks another superior uplift on the west. See Fig. 22.6. A large fresh-water lake formed on either side of the early Uinta uplift in which the Green River formation was deposited under quiet conditions and at about 1000 feet above sea level. The principal and sharp Uinta uplift then occurred with high-angle border faults, and the uplift was recorded by the deposition of the overlapping Uinta, Duchesne River, and Bridger clastics.

The Knight conglomerate in the Salt Lake–Evanston area was then cast into broad folds extending north-south, as shown in section C–C′ of Fig. 22.14.

CENTRAL UTAH

The stratigraphic relations of central Utah as worked out by Spieker (1946) are idealized in Fig. 22.17. This is the physiographic region of High Plateaus of Utah, described colorfully 80 years ago by Dutton. The formations are flat-lying and conformable but faulted on the eastern flank of the Wasatch Plateau (upper cross section of Fig. 22.18) which stands above the Colorado Plateau to the east, but on the west flank and adjacent San Pete Valley the structural relations are very complex. A monoclinal flexure is prominent on the west slope and is interpreted by Spieker as having been originally a truncated anticline following the deposition of the Price River conglomerates and the North Horn formation. The Flagstaff limestone was then deposited over the truncated anticline and then flexed downward to the west at a later time. Unconformities exist at the base of the Price River, North Horn, Flagstaff, Colton, Green River, and younger formations, and Spieker and students have postulated numerous orogenies extending from the Colorado epoch through Paleocene, Eocene, and later time. Stokes (1952), however, has pointed out the similarity between this structural complex and the salt anticlines in the Colorado Plateau, and contends that the thick Arapien shale of Jurassic age with its salt and gypsum beds has moved upward in an anticlinal core, has suffered extensive erosion, has permitted overlying beds to sag or collapse, and possibly has moved upward again on several occasions. The localized, numerous unconformities are thus explained. Hardy (1952) has shown that the central core of Arapien shale is a tight anticline, so the shale cannot be said to have flowed upward like a viscous intrusive salt body.

Farther west the remarkable Canyon Range thrust (Chistiansen, 1952) occurs and is shown in the third from the top section of Fig. 22.18. According to Christiansen a first episode of thrusting preceded the deposition of the Indianola conglomerate. The sheet probably came from the west but its roots are not evident. A second episode followed the deposition of the conglomerate.

An extensive volcanic field was built in the central part of the High Plateaus province of Utah, after the main phases of Laramide orogeny.

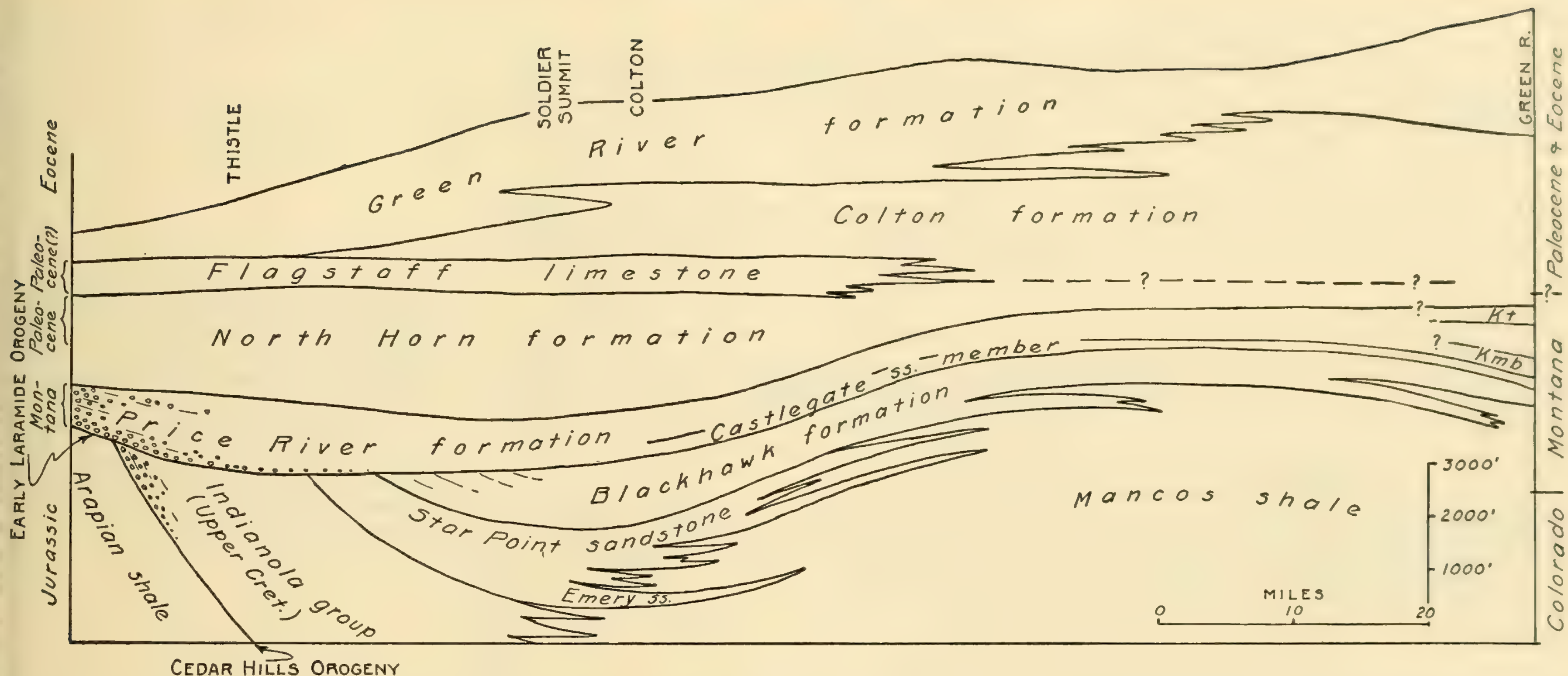


Fig. 22.17. Stratigraphic and structural relations in central Utah, in the Wasatch Plateau and adjacent areas. Section 15, Fig. 22.1. After Spieker (1946) except for the designation of the Cedar Hills orogeny. It is an idealized stratigraphic and structural diagram that extends from the Gunnison Plataau on the west to the Green River in the Colorado Plateau. Kt, Tuscher fm.; Kmb, Buck tongue.

Volcanism may have started in late Eocene time but the main eruptions appear to be Oligocene. The volcanic rocks are elaborated on in Chapter 36.

SOUTHWESTERN UTAH

The western margin of Colorado Plateau in southwestern Utah consists of a series of great steps eroded in the sedimentary rocks descending southward. These steps are cut transversely by a few long, northerly trending faults of Mid- and Late Tertiary age (Fig. 22.19). Before the faulting the western margin of the Plateau had been moderately folded, which is a transition zone into a western belt of strong folding and thrust faulting. The latter two belts are shown in Fig. 22.20. Following the folding and thrusting the Claron conglomerate of Eocene (?) age was spread over the beveled edges of the older formations. Mackin (1960) views the Claron deposition to have culminated in an extensive plain over much of southwestern Utah, upon which the later voluminous ignimbrites spread. These volcanics are discussed in Chapter 36.

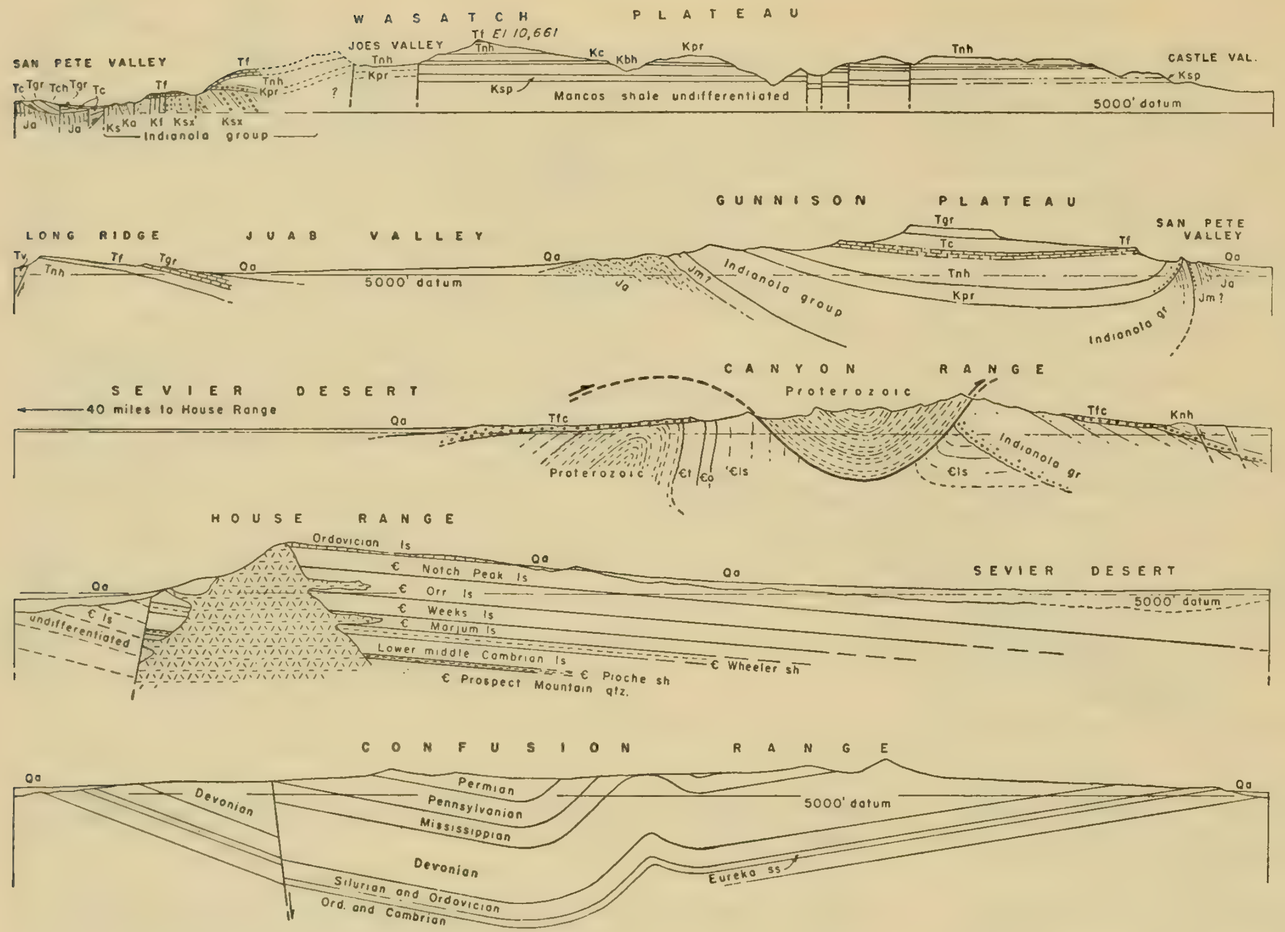


Fig. 22.18. Section from the Wasatch Plateau, westward through the Gunnison Plateau, Canyon Range (section 16, Fig. 22.1); Wasatch Plateau after Spieker, 1946; Gunnison Plateau after Spieker (1949); Canyon Range after Christiansen (1952); House and Confusion Range after unpublished map by Quigley *et al.* Tf, Flagstaff; Tnh, North Horn; Kpr, Price River; Kc, Castle Gate member; Kbh, Blackhawk; Ksp, Star Point; Ks, San Pete; Kav, Allen Valley; Kf, Funk Valley; Ksx, Sixmile Canyon; Tc, Colton; Tgr, Green River; Tch, Crazy Hollow; Tfc, Fool Creek (Oligocene; Ja, Arapien sh.; Jm, Morrison (?); €ls, Cambrian ls.; €o, Ophir sh.; €t, Tintic quartzite.

Cross sections of the Pine Valley Mountains and Hurricane Fault zone in the southwestern corner of Utah are shown in Fig. 22.21. This area is in the belt of moderate folding. We should note (1) that the porphyry pluton of the Pine Valley Mountains was intruded as a laccolith between the Claron conglomerate and the overlying ignimbrites in about mid-Tertiary time; (2) the Claron lies across a truncated fold in which the latest Cretaceous Kaiparowits formation is involved; and (3) the Hurricane fault developed after the ignimbrites were spread over the country. The evolution as conceived by Cook (1957) is represented in Fig. 22.22.

The Hurricane fault is particularly impressive because of the deep red color of some of the formations, the faulted black basalt flows, and

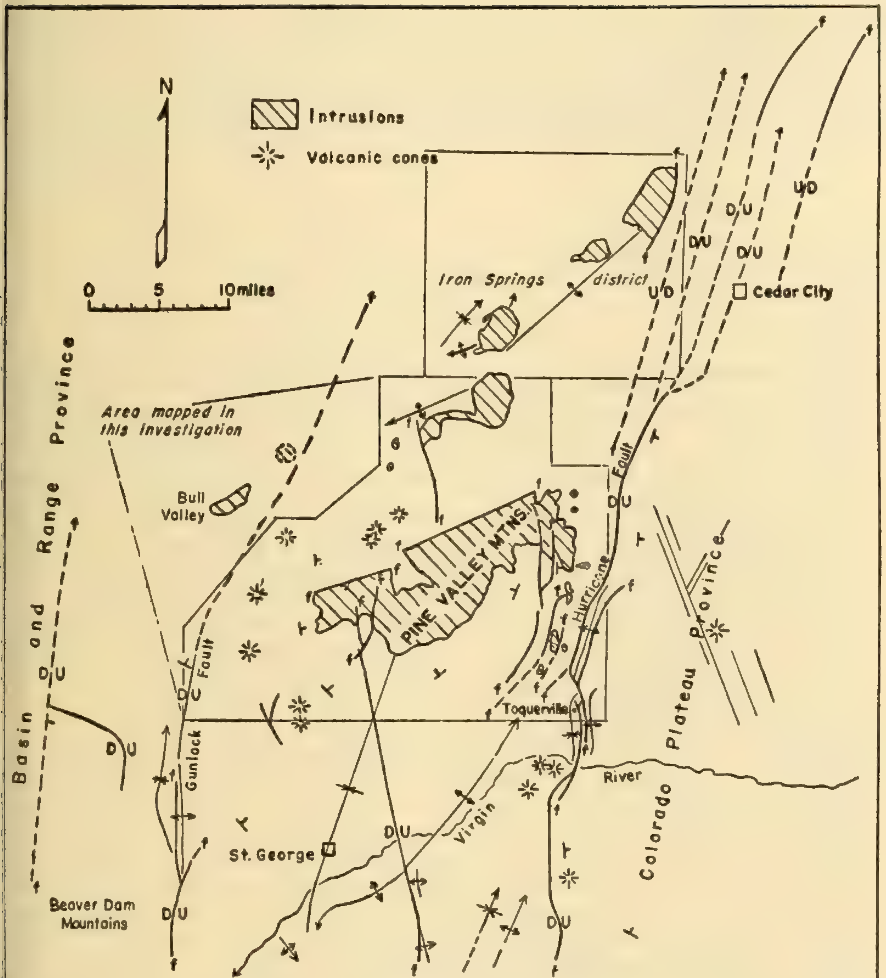


Fig. 22.19. Structural features of southwestern Utah. Reproduced from Cook, 1957.

the sparsity of soil and vegetation. The most recent displacement, so strikingly evidenced by the offset horizontal basalt remnants, represents the lesser of two periods of faulting separated by a long intererosion cycle.

In Fig. 22.19 it will be seen that two structural trends intersect. One which developed during the mid-Laramide orogeny along the boundary between shelf and geosyncline trends northeasterly. It is marked by folds, faults, and aligned intrusions. The second is a more or less north-south group of faults and belongs the Tertiary Basin and Range system. All the igneous masses in the Iron Springs district are intruded along a single horizon in the Jurassic Carmel formation (Mackin, 1947).

WESTERN UTAH

Western Utah and eastern Nevada are characterized by approximately north-south trending ranges separated by alluviated basins and generally exhibit features of Tertiary block faulting. This is the Great Basin of the geographer or the Basin and Range province of the geologist. The older internal structure of the ranges is commonly discordant with the bounding block faults, and is one of strong folding and thrusting.

In the Gold Hill district of western Utah, a complex of thrust and normal faults provides a record of prolonged orogeny. See Fig. 22.23. No accurately dated Cretaceous or Tertiary beds are present, and hence the times of orogeny are not known. Nolan (1935) believes the succession of deformational events to have spanned the Cretaceous-Tertiary boundary.

In the East Tintic Mountains of Utah (Eureka district) north-trending anticlines and synclines are superposed on a broad east-trending uplift (Morris, 1957). Overthrust faults are closely related to the folds, and cross sections (Fig. 22.24) suggest that the thrusts developed originally as bedding plane faults which later cut the beds as the folds were intensified and overturned. Some of the thrusts are themselves folded.

West of the East Tintic Mountains are the Sheeprock Mountains in which an 11,000-foot thick sequence of late Precambrian metasediments is extensively exposed (Cohenour, 1959). A tillite similar to that in the Wasatch Mountains is a prominent part. All formations—Precambrian and Paleozoic—are strongly folded and broken by thrust faults (Fig. 22.25). The major thrusts are the Sheeprock and Pole Canyon.

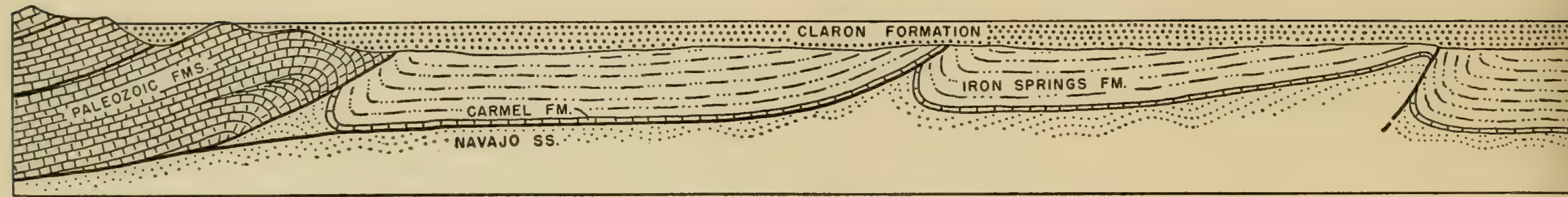


Fig. 22.20. Diagrammatic cross section of southwestern Utah restored to the time before eruptive activity started. After Mackin, 1960. Section 20, Fig. 22.11.

The complex structure is believed by Cohenour to have evolved as follows: (a) an early folding (monocline) predates the thrusting and may be of the Cedar Hills orogeny. (b) Overturning and thrusting eastward—the Sheeprock thrust formed as a wide flat sheet, now dissected so that several windows appear through it. (c) Southwesterly thrusting in which the Pole Canyon and Lion Hill thrusts formed. (d) Intrusion of the West Tintic monzonite pluton probably in Eocene time. (e) After extensive erosion basin and range faulting started and the Sheeprock granite was intruded which has been dated as Miocene. Erosion, volcanism, and renewed block faulting ensued.

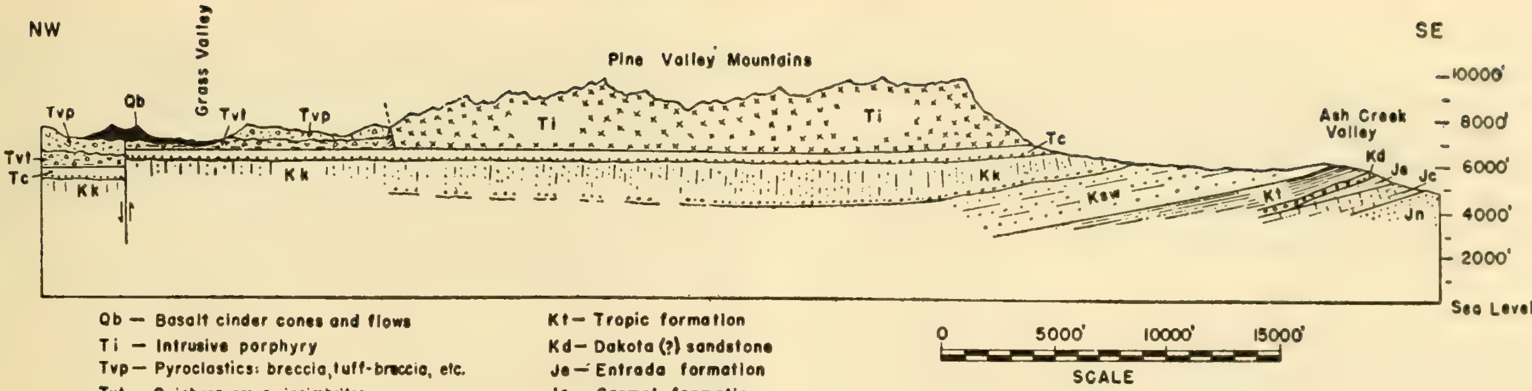
In a notable study of central-northeast Nevada and adjacent parts of Utah Misch (1960) sees eastward thrusting of the decollement type in nearly every range. He regards it as large scale. As a reasonable working hypothesis, the individually exposed decollements are considered to have moved on a regionally large plane. The main thrusting was mid-Mesozoic and predated the Laramide movements of central Utah.

SOUTHERN NEVADA

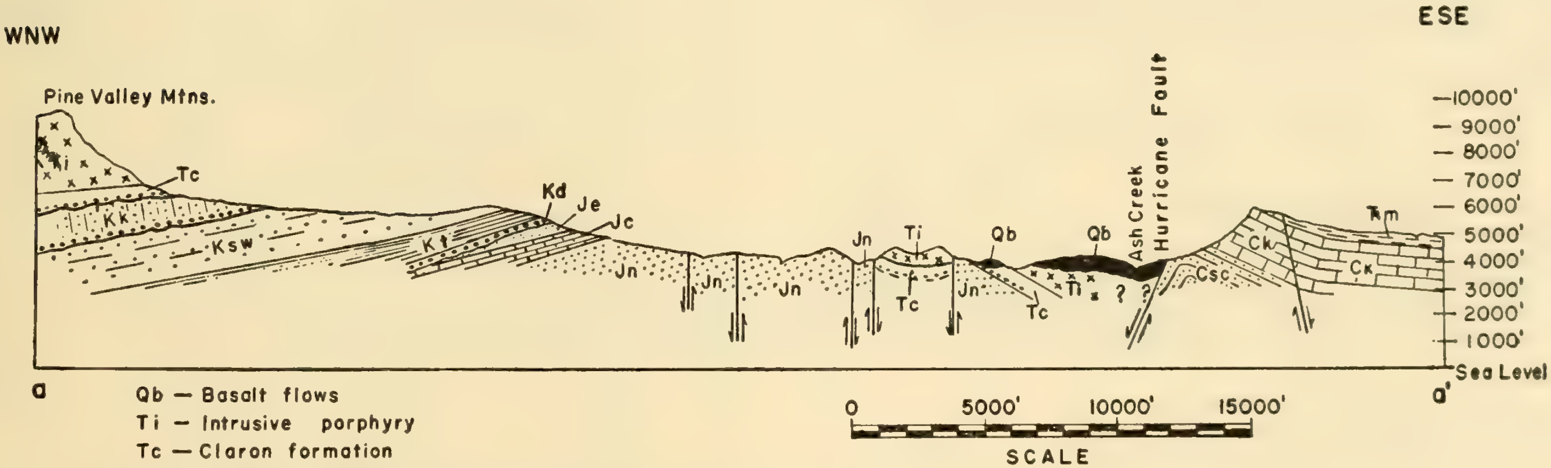
A well-known group of major thrusts occurs in southern Nevada, near the eastern margin of the Basin and Range province. Examine the *Tectonic Map of the United States.* The group comprises eight or more thrusts with variable, but commonly low, westerly dips. The absence of basement rocks in the overriding block, together with the observed changes in dip, suggests that the thrusts pass downward into a nearly horizontal sole (Hewett, 1931; Longwell, 1928). Hewett (1931) regards the thrusts in the Goodsprings area as being successively younger westward, and cites evidence indicative of an erosion interval between them. The belt of thrusting has been traced 100 miles north-south in this region.

Longwell has reported several times on the geology of southernmost Nevada, particularly on the Muddy Mountains, and his latest diagnosis of the complex structure there is as follows. Figures 22.26 and 22.27 should be referred to.

> Instead of a single large thrust in the Muddy Mountain area, Nevada, as reported from earlier field study, the writer distinguishes two superposed thrusts which may represent distinct orogenic episodes separated by a considerable time interval. Both thrusts "root" to the west. The structurally lower thrust (for which the name Muddy Mountain thrust is retained) is the more extensive; as reported previously, it has brought Paleozoic carbonate formations over Jurassic sandstone, with the heave-component of slip at least 15 miles. The higher thrust (here called the Glendale thrust) has heave-displacement of at least 5 miles. Together with associated smaller thrusts, it involves formations of early Upper Cretaceous age, as well as thick piedmont deposits that may be considerably younger. "Orogenic deposits" several thousand feet thick were laid down in front of the Glendale thrust as it advanced.
>
> Conglomerate at the base of the Upper Cretaceous section, containing boulders and cobbles derived from resistant units in older systems as low as the Permian, indicates earlier strong deformation not far west of the Muddy Mountain area. This earlier orogenic episode may have included development of the



Qb — Basalt cinder cones and flows
Ti — Intrusive porphyry
Tvp — Pyroclastics: breccia, tuff-breccia, etc.
Tvt — Quichapa group ignimbrites
Tc — Claron formation
Kk — Kaiparowits formation
Ksw — Straight Cliffs & Wahweap sandstones
Kt — Tropic formation
Kd — Dakota (?) sandstone
Je — Entrada formation
Jc — Carmel formation
Jn — Navajo sandstone



Qb — Basalt flows
Ti — Intrusive porphyry
Tc — Claron formation
Kk — Kaiparowits formation
Ksw — Straight Cliffs and Wahweap sandstones
Kt — Tropic formation
Kd — Dakota(?) sandstone
Je — Entrada formation
Jc — Carmel formation
Jn — Navajo sandstone

Ћm — Moenkopi formation
Ck — Kaibab limestone
Csc — Coconino(?) sandstone

Fig. 22.21. Sections across Pine Valley Mountains and Hurricane Fault. From Cook, 1957. Section 21, Fig. 22.1.

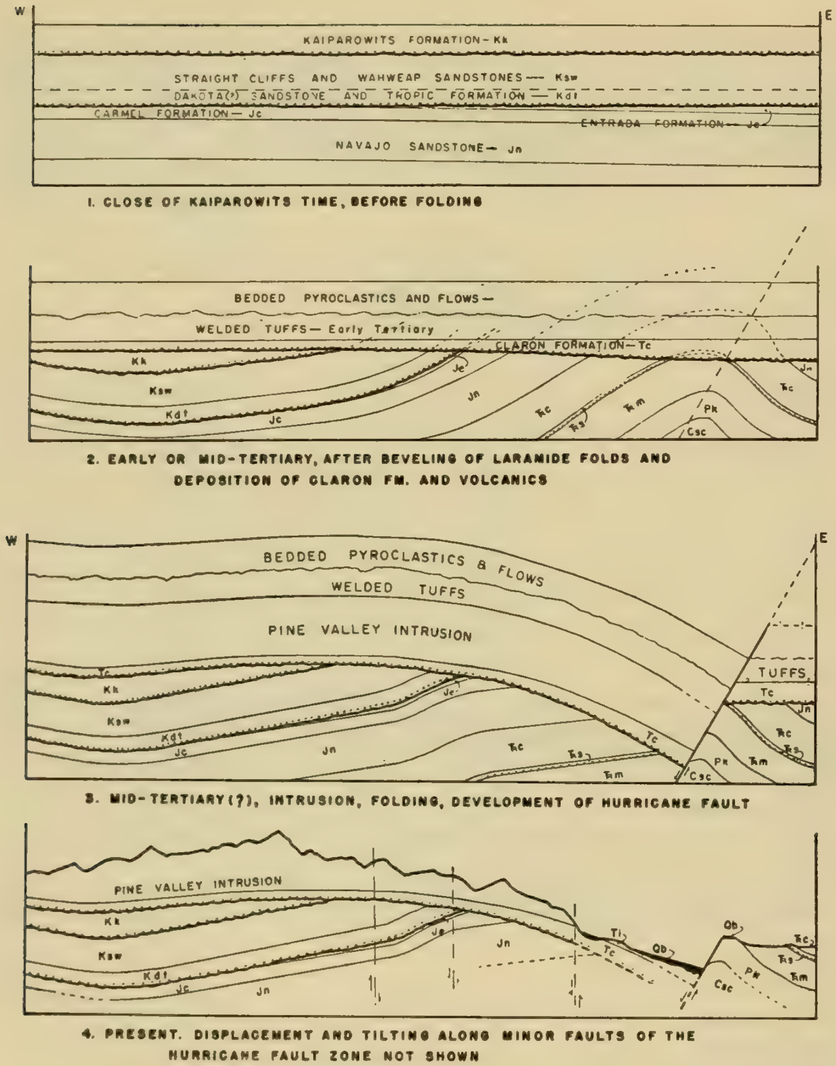


Fig. 22.22. Evolution of Pine Valley Mountains and Hurricane fault across Ash Creek Valley. Reproduced from Cook, 1957. Csc, Coconino ss.; Pk, Kaibab ls.; Ꞧm, Moenkopi fm.; Ꞧs, Shinarump congl.; Ꞧc, Chinle sh.; Jn, Navajo ss.; Jc, Carmel fm.; Je, Entrada ss.; Kdt, Dakota ss. and Tropic sh.; Ksw, Strait Cliffs and Wahweap ss.; Kaiparowits fm.; Tc, Claron fm.; Qb, Quaternary basalt.

Muddy Mountain and other large thrusts in the region which are not known to involve formations younger than Jurassic. Therefore the earlier orogeny can now be dated merely as post-Jurassic and pre-Upper Cretaceous.

Overturned and faulted folds associated with the Glendale thrust rival in complexity some structural features of the Swiss Alps. Important transverse faults with large strike-slip component pose problems of origin; the largest of these displaces the Muddy Mountain thrust plate as much as 2 miles vertically and may be genetically related to the Glendale thrust. Numerous normal faults, variously oriented, bear witness to movements ranging in date from the Glendale thrusting episode to late Cenozoic time (Longwell, 1949).

For another more recent summary treatment of southern Nevada see Longwell (1952a, b).

Fig. 22.23. Faults in the northern part of the Gold Hill Mining District, Ut. After Nolan, 1935. Section 23 of Fig. 22.1.

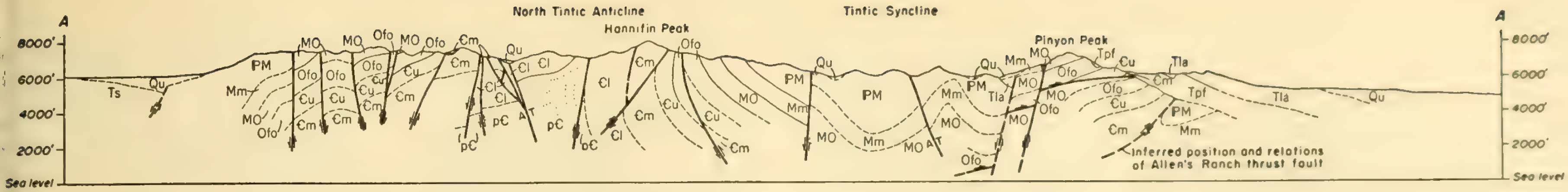


Fig. 22.24. Cross section of the East Tintic Range. By Morris, Disbrow, Lovering, and Proctor. Reproduced from Morris, 1957. Section 19, Fig. 22.1.

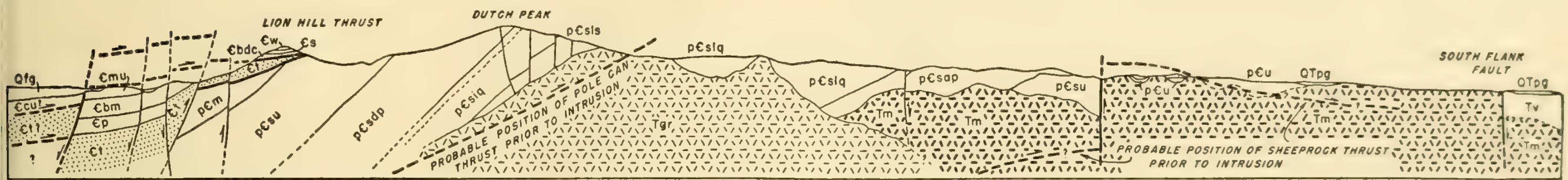


Fig. 22.25. Cross section of the Sheeprock Mountains. After Cohenour, 1959. Section 18, Fig. 22.1.

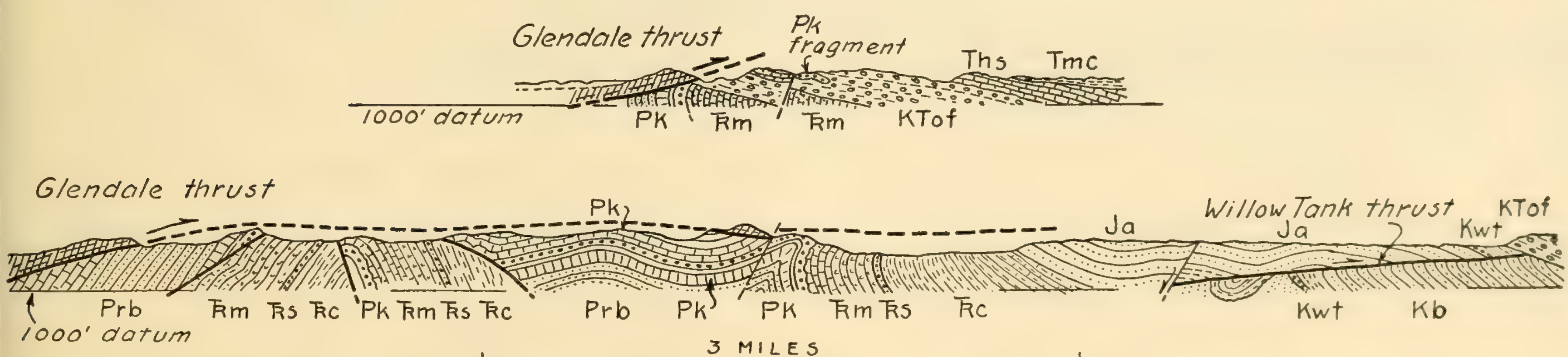


Fig. 22.26. West to east sections through northern Muddy Mountains, after Longwell, 1949, showing Glendale and Willow Tank thrust sheets. Glendale thrust sheet is made up of Cambrian to Pennsylvanian strata. Prb, Permian red beds; Pk, Permian Kaibab ls.; TRm, Moenkopi fm.; TRs, Shinarump congl.; TRc, Chinle fm.; Ja, Aztec ss.; Kwt, Willow Tank congl.; Kb, baseline ss.; Ktof, Overton fm.; Ths, Horse Spring fm.; Tmc, Miocene (?) Muddy Creek fm. Section 20 of Fig. 22.2.

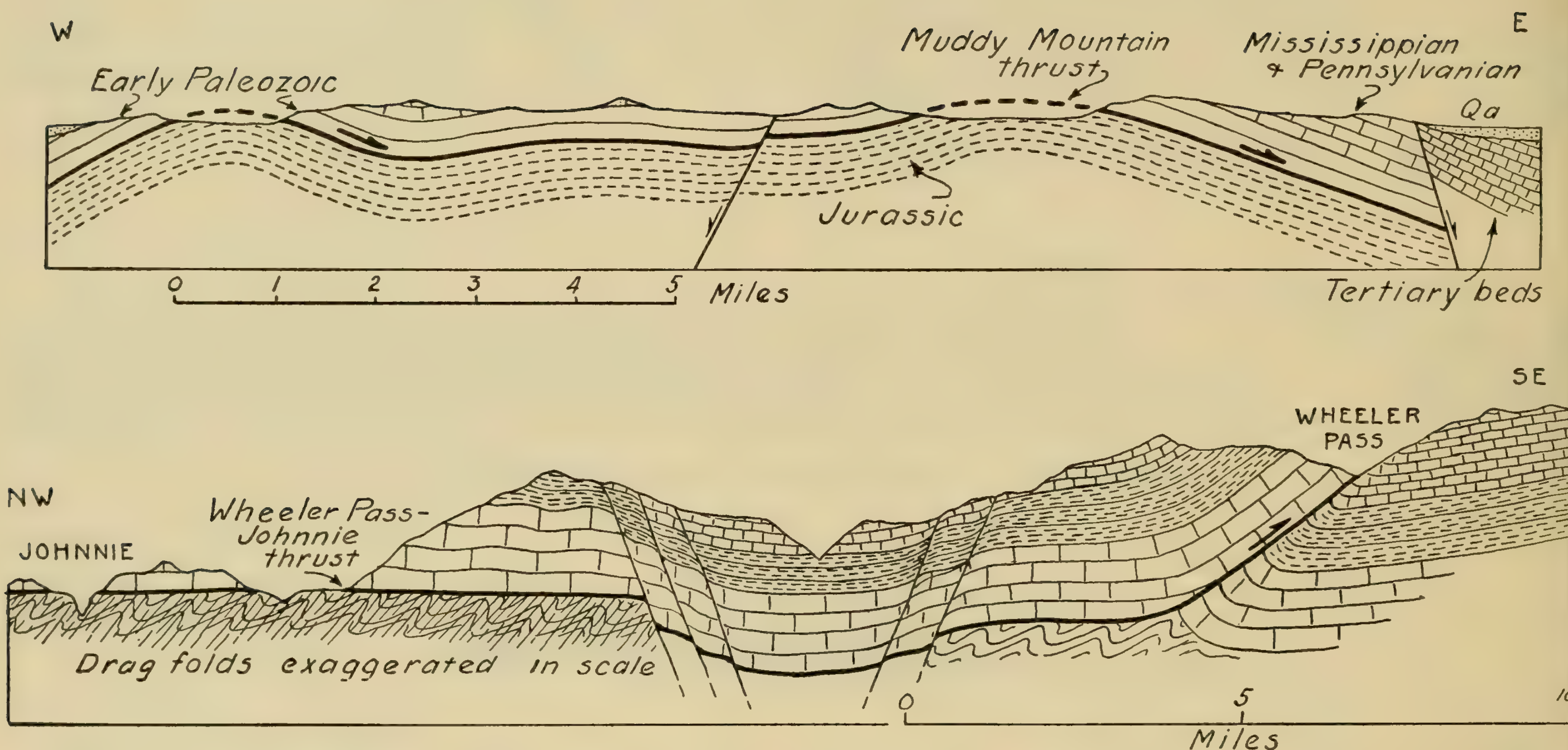


Fig. 22.27. Generalized cross sections of the Muddy Mountain thrust (upper diagram) and Wheeler Pass–Johnnie thrust (lower diagram). In the upper diagram, section 22 of Fig. 22.1, the Paleozoic strata are limestone and dolomite, and the Jurassic strata are thick-bedded sandstone. In the lower diagram about 21,000 feet of strata are shown in the thrust sheet, but 8000 more of younger strata were probably affected by the thrust. The Johnnie and Wheeler Pass area is 60 to 80 miles west of the Muddy Mountains, section 23 of Fig. 22.1. Both diagrams after Longwell, 1945.

23.

CENTRAL MONTANA ROCKIES

GENERAL FEATURES

The structures included under the name, Central Montana Rockies, are those in Montana east of the Canadian and Montana Rockies and north of the Wyoming Rockies. The boundary southeasterly of the Foothill belt (see Fig. 23.1) is not clearlly defined and is drawn chiefly for the purpose of discussion. The transition from the Cordilleran geosyncline to the shelf is approximately along the west side of the Foothill belt and along the east side of the central Rockies, so that the central Montana Rockies are developed from the shelf. In Chapter 5 it will be recalled that the Big Snowy basin formed in an east-west direction through central Montana in Mississippian and early Pennsylvanian time and merged into the Williston basin in eastern Montana, the Dakotas and southern Saskatchewan and Manitoba. Also an arm of the Beltian basin extended eastward through the Little Belt Mountains.

In the Laramide orogeny a major zone of domes and monoclinal flexures formed in an east-west direction approximately in the site of the older Big Snowy basin, and in addition, six smaller, subcircular mountain groups evolved about igneous centers. Some striking *en echelon* fault zones also developed. The mountain groups rise imposingly 2000 to 5000 feet from the plains. The primary cause of the entire assemblage of mountain groups is probably magmatic. Major intrusions into the Precambrian rocks domed up the Paleozoic and Mesozoic veneer, and in places central vents and associated dike swarms broke through to the surface or fed sill and laccolithic intrusions into the Cretaceous strata a short distance below the surface. A certain amount of horizontally acting crustal stress was relieved at about the same time as the intrusions. This is attested by the *en echelon* fault zones. The magmatic theory is elaborated upon in Chapters 19 and 36 which discuss the igneous provinces of the western United States.

CENTRAL ZONE OF UPLIFTS

The central zone of uplifts extends from the Little Belt Mountains on the west to the Porcupine dome on the east. The Big Snowy Mountains comprise a prominent and perhaps the best-known dome. See Figs. 23.2 and 23.3. The Madison limestone has proved very resistent to erosion and forms the surface rock of a large central area. The Big Snowy dome has a length of about 25 miles and a width of 12 miles; the structural relief is 12,000 feet on the steep southern flank.

The dome of the Little Belt Mountains is much larger in horizontal dimensions than the Big Snowy, but not so complete. Its southwestern half is composed of Beltian strata intruded by granodioritic batholiths and the domal structure obliterated by sharp folds and thrust faults. Its northeasterly flank is gentle and stretches beyond the Highwood volcanic group to the Blood Creek syncline with a structural relief of 11,000 feet.

Fig. 23.1. Mountain basins and uplifts of Montana. See Fig. 23.2 for structural details of central Montana. The Sweetgrass arch includes the Kevin-Sunburst dome and the South arch. The Genou trend is a relief feature of the Precambrian surface which was buried by onlapping Cambrian and Devonian strata (Alpha, 1955).

Fig. 23.2. Structure contour map of central Montana. Reproduced from Reeves, 1930.

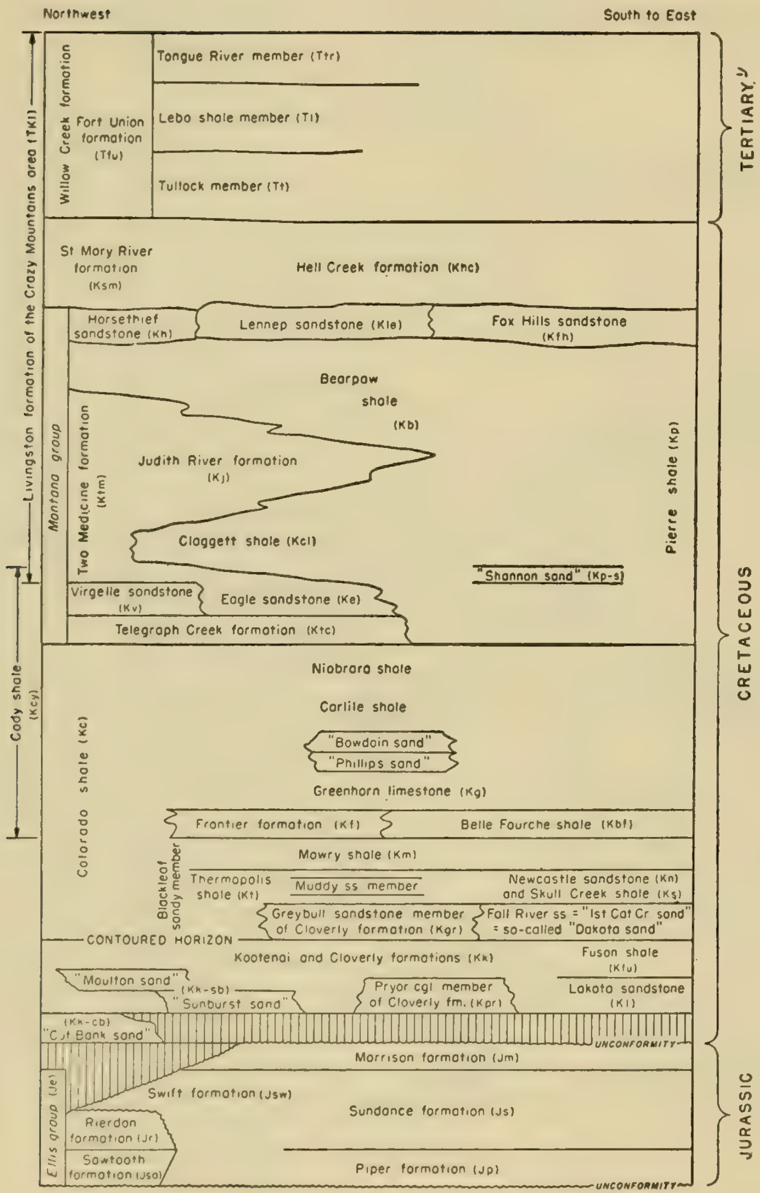


Fig. 23.3. Stratigraphic diagram of Cretaceous and Paleocene formations of Montana. Reproduced from Dobbin and Erdman, 1955.

Most of the anticlines and domes of the central zone of uplifts are asymmetric and can be considered as flexures. The Cat Creek anticline which extends eastward from the Judith Mountains, is a good example. Thom (1923) considers the flexures due to draping of the flexible surficial strata over faulted blocks of the more brittle Precambrian rocks below. See lower diagram of Fig. 23.3.

Central and eastern Montana and northern Wyoming are the sites of a remarkable succession of formations that bridge the Cretaceous and Tertiary periods. The position of the boundary of the two stratigraphic systems has been a matter of lively argument and study for many years. The chart of Fig. 23.4 shows the age assignments of the various formations on the *Structure Contour Map of the Montana Plains* by Dobbin and Erdman (1955). The youngest formation generally reported upturned in the flexing is the Fort Union, which is regarded as Paleocene. It is known that the Big Horn Mountains arch had already risen and been stripped to the Precambrian by Fort Union time, and that further uplift occurred soon afterward. Since the Big Horn arch extends into south-central Montana, in the proximity of the east-west flexures and related domes just to the north, it is possible that the central Montana structures came into existence during the same period, viz., immediate pre-Fort Union and again in post-Fort Union.

The laccoliths, dikes, and stocks are nearly all in Cretaceous strata; only in the Bearpaw Mountains has a deposit as young as the Teritary in association with the volcanics been reported. The oldest volcanics there rest mostly on the Fort Union formation, but in places a cobble layer intervenes which may be Eocene or even Oligocene in age. At least the oldest extrusions, which are older than the stocks of the area, are younger than the cobble layer, and therefore at least lower Eocene in age, and perhaps younger (Pecora, 1941).

ZONES OF EN ECHELON FAULTS

Characteristics and Structural Relations

Three zones of *en echelon* faults occur in central and south-central Montana. The Cat Creek fault zone is at the north, the Lake basin fault zone in the middle, and the Nye–Bowler zone on the south, just north of

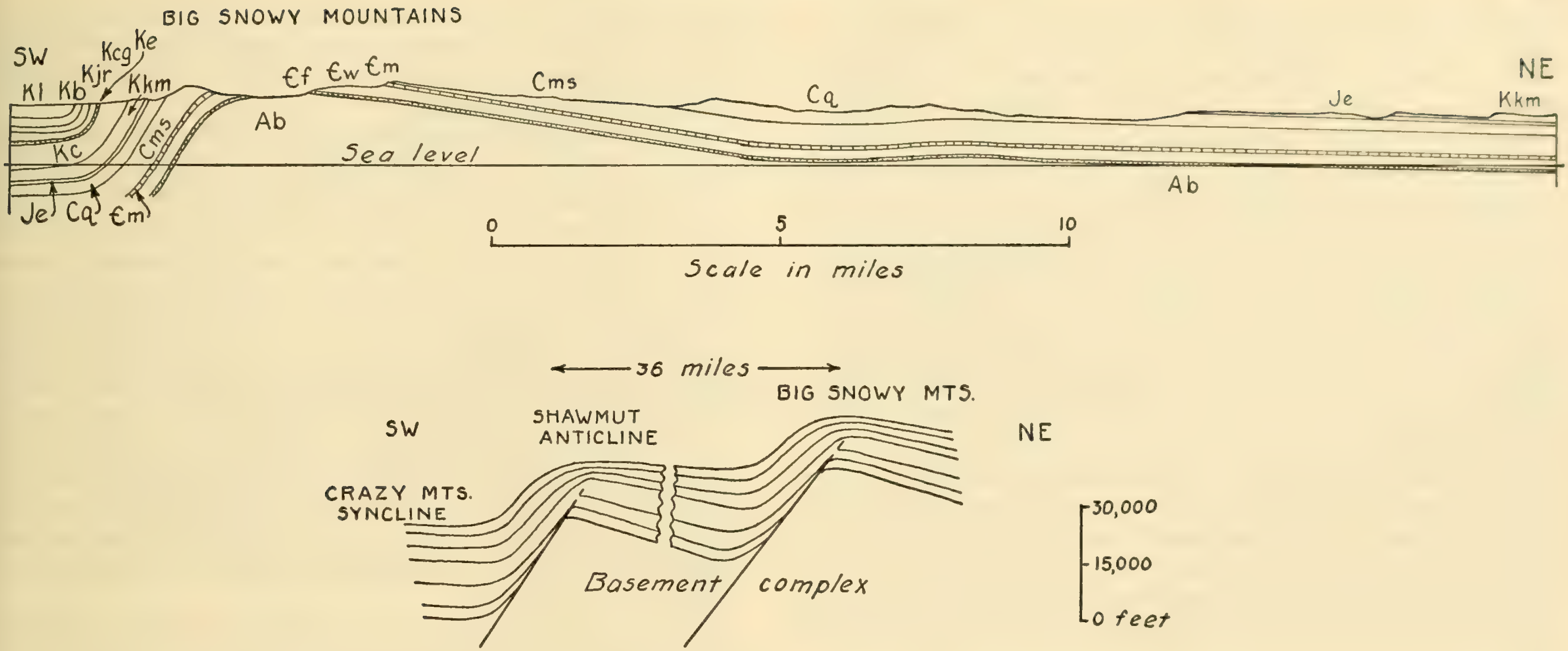


Fig. 23.4. Upper diagram is section across the Big Snowy Mountains, after Reeves, 1931. Lower diagram illustrates the relation of flexures in the surficial sedimentary rocks to deep-seated faults in the Musselshell Valley region of Montana, after Thom, 1923.

the Beartooth uplift. See Figs. 23.1 and 23.2. The Cat Creek anticline (monoclinal flexure) extends eastward from the Judith Mountains, and the strata of the steep flank are broken by a series of small faults. It appears that the north block, the Blood Creek syncline, moved slightly westward in the downward movement.

The Lake basin fault zone extends east and west of Billings, is the largest of the three, and has a length of over 100 miles. On the west end it cuts the south flank of the Big Coulee–Hailstone dome, and on the east end it cuts the northward dipping strata from the Big Horn uplift. For the most part, the southeast side of each fault is downthrown, but there are many exceptions (Hancock, 1918). Along some of the faults, the direction of throw changes from one end to the other. In any event, the throw is small. The fault planes are generally inclined 30 to 80 degrees. It would seem that the zone of faults came into existence after the dome and flexure that it cuts and that a deep-seated fault with horizontal movement was the cause. The surficial strata over the horizontally displaced blocks broke in numerous small tensional faults oriented obliquely to the master fault beneath (Chamberlin, 1919). As with the Cat Creek zone, the north block moved westward.

The Nye–Bowler zone consists of a series of anticlines, domes and half domes in perfect alignment for 56 miles, extending from the Beartooth Mountain front to the Pryor Mountains. Dips on the south limb are

steeper than on the north. Cutting the anticlines are two principle sets of faults, viz., faults parallel with the strike of the zone on the anticlinal axes, and faults in *en echelon* arrangement diagonally across the zone. There are also a number of feather faults that terminate at the longitudinal faults, together with two downdropped fault blocks. It has also been noted that the formations vary in thickness across the zone or "lineament," as Wilson (1936) calls it. The Lance formation and Lebo member of the Fort Union thicken abruptly south of the axis of the fold. This is taken to mean the beginning of flexing or faulting during the deposition of the uppermost Cretaceous and Paleocene beds.

Centers of volcanism are also aligned with the Nye–Bowler lineament. The laccoliths of Limestone Butte and Round Mountain were intruded along the westward projection of the belt. Also, the intrusions of Green Mountain and Squaw Peak in the McLeod area came up along the lineament (Wilson, 1936).

The assemblage and relation of all the structural features of the Nye–Bowler lineament have led Wilson (1936) to conclude that they are the surface expression of a single deep-seated fault, along which both vertical and horizontal movement took place. The south block both sank and moved eastward. This horizontal movement is in the same direction as that of the blocks north and south of the Lake basin and Cat Creek fault zones.

STAGES OF OROGENY

The first perceptible stage of the Laramide orogeny in south-central Montana was marked by slightly coarser sediments in the Judith River formation and by the eruption of volcanoes that may have heralded the early rise of the Beartooth block. The earliest movement on the basement fault of the Nye–Bowler lineament also occurred, and dikes were intruded and agglomerate (the lower part of the Livingston formation) piled up along the fault trace. The second stage, according to Wilson, lasted through the deposition of the Bearpaw, Lennep, Colgate, and Lance formations and the Lebo member of the Fort Union. The Nye–Bowler flexure first appeared in the area of deposition of the Upper Cretaceous sediments just mentioned, and they accumulated thinly over the crest and thickly on the depressed area to the south. The *en echelon* faults of normal displacement also were formed at this time. Volcanism continued from the first stage all through the second.

The third stage was principally that of uplift of the Beartooth Mountains, and with the orogeny the Nye–Bowler monocline was compressed and arched, with the formation of its individual domes. Its normal faults in part became reverse ones in face of the compression; more horizontal movement occurred, and the feather faults came into existence. Erosion was actively attacking the rising Beartooths, and the waste products were spread out to the east as the Tongue River sandstones and shales. By the time of the next uplift of the Beartooth block, which would be the fourth stage, the Precambrian crystalline rocks in it had been exposed and the beds of flexure had been considerably truncated. The following wave of debris from the Beartooths spread over the truncated structures of the basin. All this deposit, the "Wasatch sandstones and conglomerates," has subsequently been eroded away save for a downfaulted and protected block. The block faulting marks the last and fifth stage of the Laramide orogeny in the region. This last stage should possibly be considered post-Laramide.

According to a recent study by McMannis (1958) three major spreadings of andesitic debris into the Crazy Mountain basin from the southwest occurred, each one reaching further than the other. These were in Judith River, Lennep, and Lebo times. The Lennep and Lebo "pulses" of McMannis make up the second stage of Wilson above reviewed. The Lebo volcanism and uplift constituted the culmination of the Laramide orogeny, according to McMannis, and this occurred in Paleocene time.

IGNEOUS CENTERS

Distribution and General Structure

Six igneous mountain clusters of the Central Montana Rockies may be recognized as follows, beginning on the northwest: the Sweetgrass Hills, the Bearpaw Mountains, the Little Rocky Mountains, the Highwood Mountains (Fig. 23.1), the Moccasin and Judith Mountains (Fig. 23.2),

and the Crazy Mountains. Those where local domes have been created, presumably by laccolithic intrusions in the Cretaceous strata, are Sweetgrass Hills, Bearpaw, Little Rocky, Moccasin and Judith Mountains. The other two groups, the Highwood and Crazy Mountains, are characterized by remarkable radiate dike swarms and not by domal uplifts. The Highwood Mountains are on the north flank of the Little Belt Mountains, and the Crazy Mountains are directly in the lowest part of the Crazy Mountains basin *(Structure Contour Map of the Montana Plains).*

Bearpaw Mountains

The Bearpaw Mountains are made up of two large volcanic fields with a central strip, 2 to 8 miles wide, of deformed and metamorphosed sedimentary rocks, known as the Bearpaw Mountains structural arch. It trends N 60° to 80° E as does an accompanying swarm of thousands of dikes (Pecora, 1957). The oldest formation involved is the Madison, and the youngest the Judith River of Late Cretaceous age.

The arch was first developed as a prevolcanic structure and continued to develop throughout the magmatic history. Vertical uplift of 5000 to 7000 feet is demonstrable, with block faulting in prevolcanic time permitting a good part of the uplift in places.

The great abundance of Precambrian basement inclusions in the rocks of latitic composition represents transportation vertically of at least 2 miles through the Paleozoic and younger formations and at least 4 miles if the volcanic pile is also pierced (which is 10,000 to 15,000 feet in maximum thickness). The extensive distribution of the inclusion-bearing felsic rocks over 1600 square miles of the Bearpaw Mountain uplift area and the absence of quartzite fragments representative of the Belt series are significant relationships that may indicate either an angular uncomformity and the removal of the late Precambrian rocks of the Belt series before deposition in the early Paleozoic sea in this region or a development of the felsic magma very deep in the basement itself (Pecora, 1957).

The volcanic activity ran its course during middle and late Eocene, and radiogenic ages of zircons in a syenite are reported to be about 40 to 60 m.y. Post-volcanic faulting and intrusions have disturbed the original attitude of much of the layering of the volcanic pile.

A great variety of mafic subsilicic-alkalic to felsic silicic-alkalic rocks occur, with the mafic rocks exceeding the felsic in volume.

An extensive skirt of small thrust faults flanks on the south the Bearpaw igneous centers and domal uplift, and is regarded by Reeves (1946) as a gravity slide phenomenon down slope from the uplift.

Little Rocky Mountains

The Little Rocky Mountains are a singular structural type. They lie apart—somewhat north—of the other uplifts and domes of central Montana, and are erosional features of a subcircular dome, about 20 miles in diameter, which embraces more than 50 faulted subordinate domes. See Fig. 23.5. Alkalic igneous rocks of Tertiary age, mainly in the form of sills, have intruded the Cambrian strata but are not known to have intruded sedimentary rocks younger than Cambrian. All their contacts with post-Cambrian rocks appear to be fault contacts, and indicate that the igneous rocks, after having consolidated at and near the base of the Cambrian, were deformed, broken, and faulted by upward pressure, probably due to an underlying, rising magma (Knechtel, 1944).

The subordinate domes on the large subcircular dome of the Little Rocky Mountains were formed by bodies of igneous rock which were punched upward into the sedimentary rocks. They range in diameter from 1½ to 3½ miles. Each is typically subcircular or subelliptical in plan and normally includes a hinged block that has been raised like a trap door (Knechtel, 1944). See cross section of Fig. 23.5.

Relation of Igneous Activity to Tilted Fault Blocks

The Little Rocky, Moccasin, and Judith Mountains, the domes of the Big Belt, and those southeast of the Little Belt Mountains, and possibly the Porcupine dome are variations of laccolithic uplifts. The Highwood and Crazy Mountains are the products largely of extrusive activity, but stocks, numerous dikes, and laccoliths are present.

In spite of the flexures and deep-seated faults beneath them, the sedimentary beds were almost horizontal in most places at the time of igneous activity. The belt of vigorous Laramide deformation lay to the west. The laccolithic intrusions especially domed up the beds, but in the two mountain groups where extrusive rocks are most abundant, the strata seem little deformed below the volcanics.

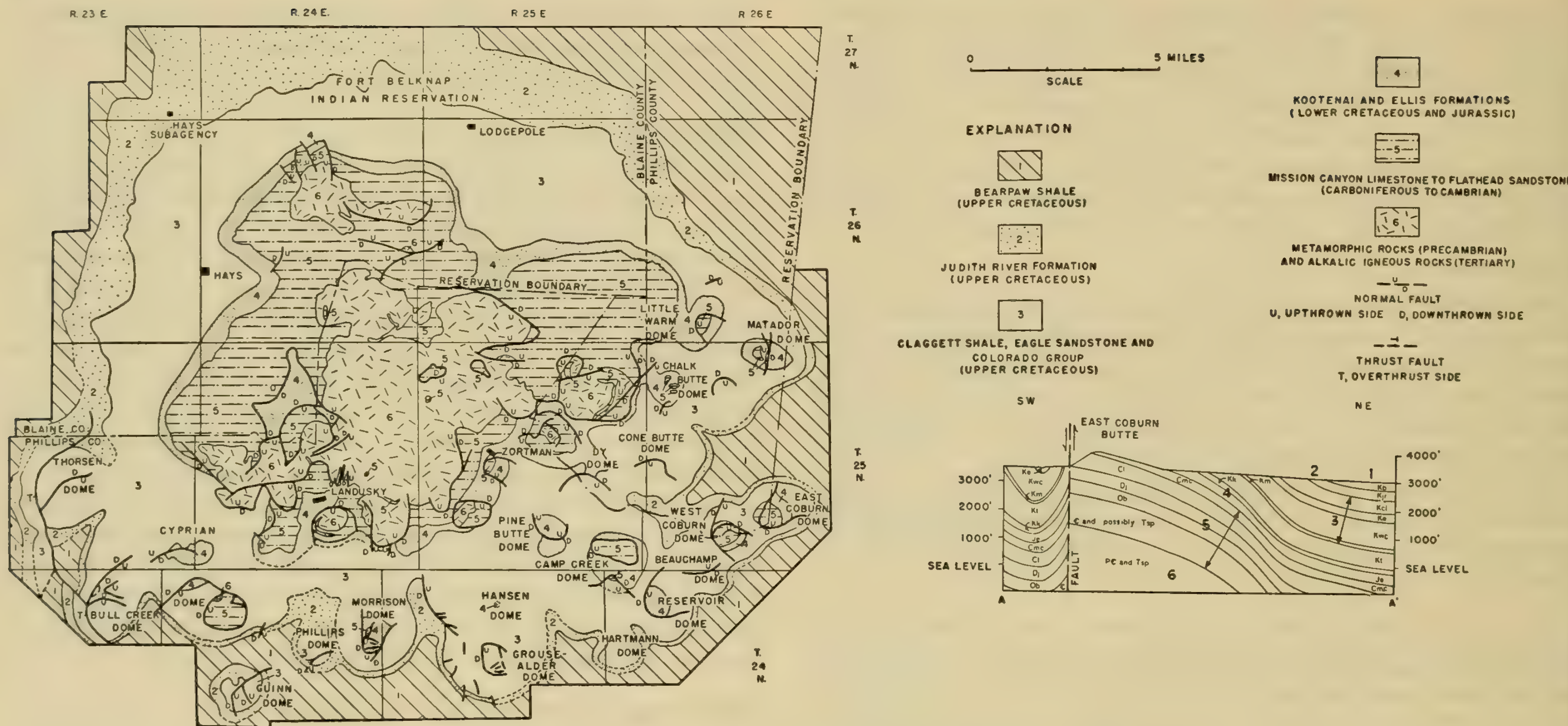


Fig. 23.5. Geologic map of Little Rocky Mountains, Montana. Reproduced from Knechtel, 1944.

An approximate parallelism of the volcanic groups with the major faults and flexures of central Montana has been pointed out by Thom (1923), but it is evident from inspection of the *Tectonic Map of the United States* that the major faults are clearly not the loci of the magmatic activity. However, the subparallel alignment and contemporaneity of origin lead Thom to view all the faults, flexures, and igneous rocks as tied to the regional deforming forces of the Laramide orogeny.

Petrology

The igneous rocks range from rhyolites to basalts in one category and from shonkinites through nepheline syenites to syenites—rocks that are rich in potash and soda and almost devoid of plagioclase—in another. The rocks of each mountain group fall into one or more eruptive stages; and the rocks of each stage have peculiar mineral and chemical features, although they commonly range from highly mafic to highly felsic. Each

stage is separated from the other by intervals during which few or no eruptions occurred, but instead, extensive erosion. Chapter 33 deals with the origin of the igneous rocks in this province and should be referred to for a discussion of the igneous and tectonic provinces of the western United States.

In each of the stages a rock near the mafic end is believed to represent the primary magma. This rock ranges from an ordinary basalt to orthoclase basalt to plagioclase shonkinite to shonkinite rich in potash and lacking plagioclase. The gradational character of the eruptive stages and their close association in time and space indicate a common origin (Larsen, 1940). Two periods of magmatic differentiation are required: first, a deep-seated differentiation to yield the primary magmas of the individual eruptive stages, and second, a shallower differentiation of the primary magmas which were probably derived from a basaltic magma by the removal of crystals of calcic plagioclase and hypersthene in depth. The relative flatness of the sedimentary rocks into which and through which the magmas have moved indicates that the magmas have not been disturbed by orogenic forces; therefore they could have differentiated during the long, quiet interval which seems necessary. The second period of magmatic differentiation by crystal settling was characterized, in most stages, by assimilation of siliceous material. The amount of assimilated material was especially large in the Crazy and Little Belt Mountains where syenites were followed by granites.

The Shonkin Sag laccolith, one of nine in the Highwood Mountains, is worth special mention. It has long been held as a classic example of magmatic differentiation in place, but the theory has been questioned and one of multiple intrusions proposed (Barksdale, 1937). More recently, Hurlbut and Griggs (1939) contend that the first theory has the greatest merit. In describing the laccoliths of the Highwood Mountains they point out, first, that they are broad, sill-like bodies and not the domed-shaped ones that Gilbert (1877) pictured in the Henry Mountains of Utah, and second, that the peripheral contacts are not simple wedges of intrusive rock, but a complex of multiple sills, crumpled strata, and small normal and reverse faults. Examine Fig. 23.6.

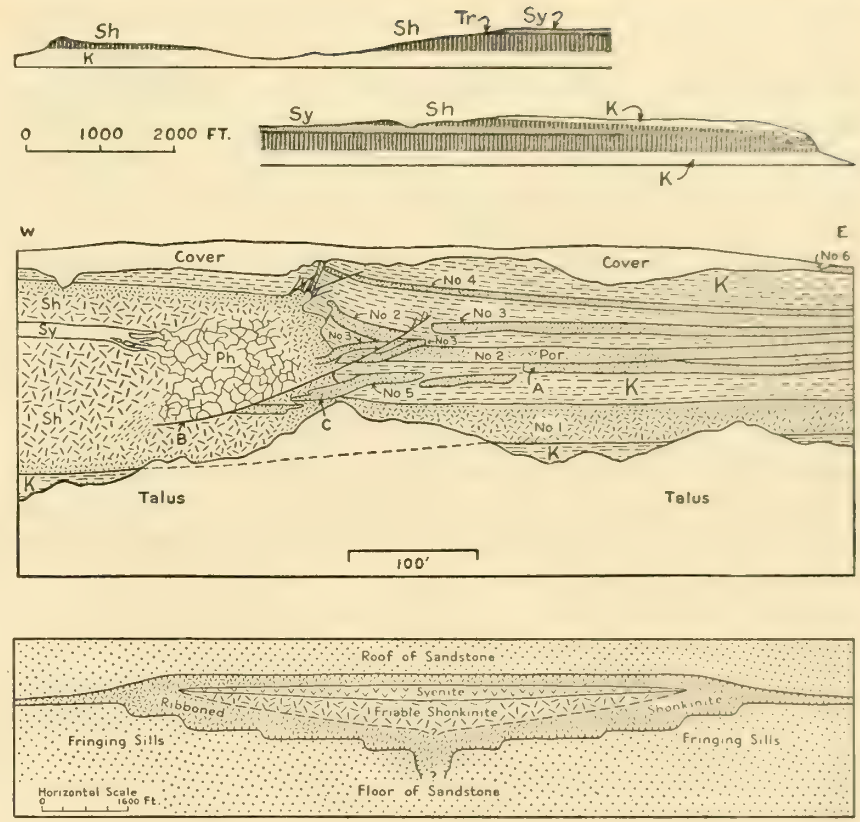


Fig. 23.6. Upper section: the Shonkin Sag laccolith, K is Cretaceous sandstone, Sh is shonkinite, Tr is transition rock, and Sy is syenite. After Hurlbut, 1939.
Middle section: detail of eastern termination of Shonkin Sag laccolith, K, Cretaceous strata, Sh is shonkinite, Sy is syenite, Ph is phonolite, Nos. 1 to 5 are sills of shonkinite porphyry. After Hurlbut, 1939.
Lower section: diagrammatic section of Boxelder laccolith. After Pecora, 1941.

The main body of the Shonkin Sag laccolith is made up of three horizontal layers, an upper one and a lower one of shonkinite, and an intermediate one of syenite. This is true of all the laccoliths in the group; the larger the pluton, the greater the amount of syenite. According to the

theory of separate injections, the syenite magma was injected into a partially solidified shonkinite; but according to the theory of magmatic differentiation in place, the syenite was formed by the settling of heavy minerals out of the shonkinite magma and the rising of leucite crystals. The minor injections of the shonkinite in the syenite at the lower contact of the syenite are explained as due to surges of magma incident to deformation of the magma chamber.

The Boxelder laccolith of the Bearpaw Mountains is also an instructive example of differentiation in place (Pecora, 1941) and the lower cross section of Fig. 23.6 has been prepared to show the relations.

STRUCTURES OF THE NORTHERN GREAT PLAINS

East of the zone of flexures and domes and north of the Black Hills is a long, asymmetrical, gentle fold known both as the Cedar Creek anticline and the Baker–Glendive anticline. See Fig. 23.1. Between it and the Porcupine dome is the shallow Sheep Mountain syncline. All are Laramide structures. They are so gentle, however, that they hardly deserve inclusion in any belt of Laramide orogeny. The very low Bowdoin dome northeast of the Bearpaw Mountains is in the same class. The Cedar Creek anticline has produced commercial gas from the Upper Cretaceous strata in several local domes along it, and deep wells have shown the presence of the Lower and Upper Mississippian strata there, and consequently the extension eastward of the Big Snowy trough (De Wolf and West, 1939). One reached the Precambrian at a depth of 9680 feet, having passed through 3920 feet of Upper Cretaceous strata, 220 feet of Lower Cretaceous, 1450 feet of Jurassic and Triassic, and 4090 feet of Poleozoic (Seager, 1942). Oil was found in a local dome, the Pine field, on the anticline in 1952 in Ordovician and Silurian strata. Several other small anticlines and domes in the setting of the major structures previously described, have been drilled and produce oil. The Charles evaporite sequence is a prominent productive zone.

24.

WYOMING ROCKIES

GENERAL CHARACTERISTICS

The topographic features of Wyoming are for the most part large, northwest-trending ranges and large intermontane basins. Study Fig. 24.1. Of the ranges, the most imposing are the Beartooth, Absaroka, Wind River, and Big Horn. Numerous peaks in these ranges reach elevations above 12,000 feet and stand 5000 to 7000 feet above the basin floors. Other ranges, now not so high and partly buried by Tertiary sediments, were undoubtedly once very high and are equally important structural elements. The Wyoming structural system is defined for convenience as extending slightly beyond the borders of the state. The Pryor Mountains at the north end of the Big Horn and the Beartooth Range extend into southern Montana; the Black Hills lie mostly in western South Dakota, and the Uinta Range mostly in Utah. On the other hand, the Colorado Rockies extend into southeastern Wyoming by way of the Laramie, Medicine Bow, and Park ranges. Certainly the Colorado and Wyoming rockies are closely related, and any separation structurally is arbitrary and for the sake of organization.

The Wyoming Rockies have been referred to as the outer ranges or shelf ranges of the Rocky Mountain Cordillera, in contrast to the inner or geosynclinal. This point has been discussed in the introduction to the general subject of the Late Cretaceous and Early Tertiary Rocky Mountain systems, Chapter 19. By inspection of the paleotectonic maps of the Paleozoic and Mesozoic eras, it will be apparent that the area of the outer ranges was generally one of shelf seas except in Late Cretaceous time, when in certain basins of Wyoming and Colorado over 10,000 feet of strata accumulated.

In addition to a rather thin veneer of Paleozoic, Triassic, and Jurassic sediments the ranges have extensive, oval-shaped cores of Precambrian rock, and for the most part are asymmetrical uplifts either in the form of large anticlines or great tilted fault blocks. The Absaroka Range is an exception because it is composed chiefly of pyroclastics and volcanic flows of a date later than most of the other mountain building. The Absarokas are connected with and closely related to the volcanic plateau of Yellowstone Park.

TETON—GROS VENTRE—WIND RIVER ELEMENT

The Teton, Gros Ventre, and Wind River ranges are in general alignment and extend from the Idaho line south of Yellowstone Park southeastward for 150 miles. They are of great height and beauty, and support a number of small glaciers. The Grand Teton is 13,747 feet high, and Gannett Peak in the Wind River Range is 13,785 feet high. These are the highest peaks in Wyoming. All three ranges have Precambrian crystalline cores and fairly simple structure along their northeastern flank, such as characterizes the great anticlinal arches of the Big Horn and Black

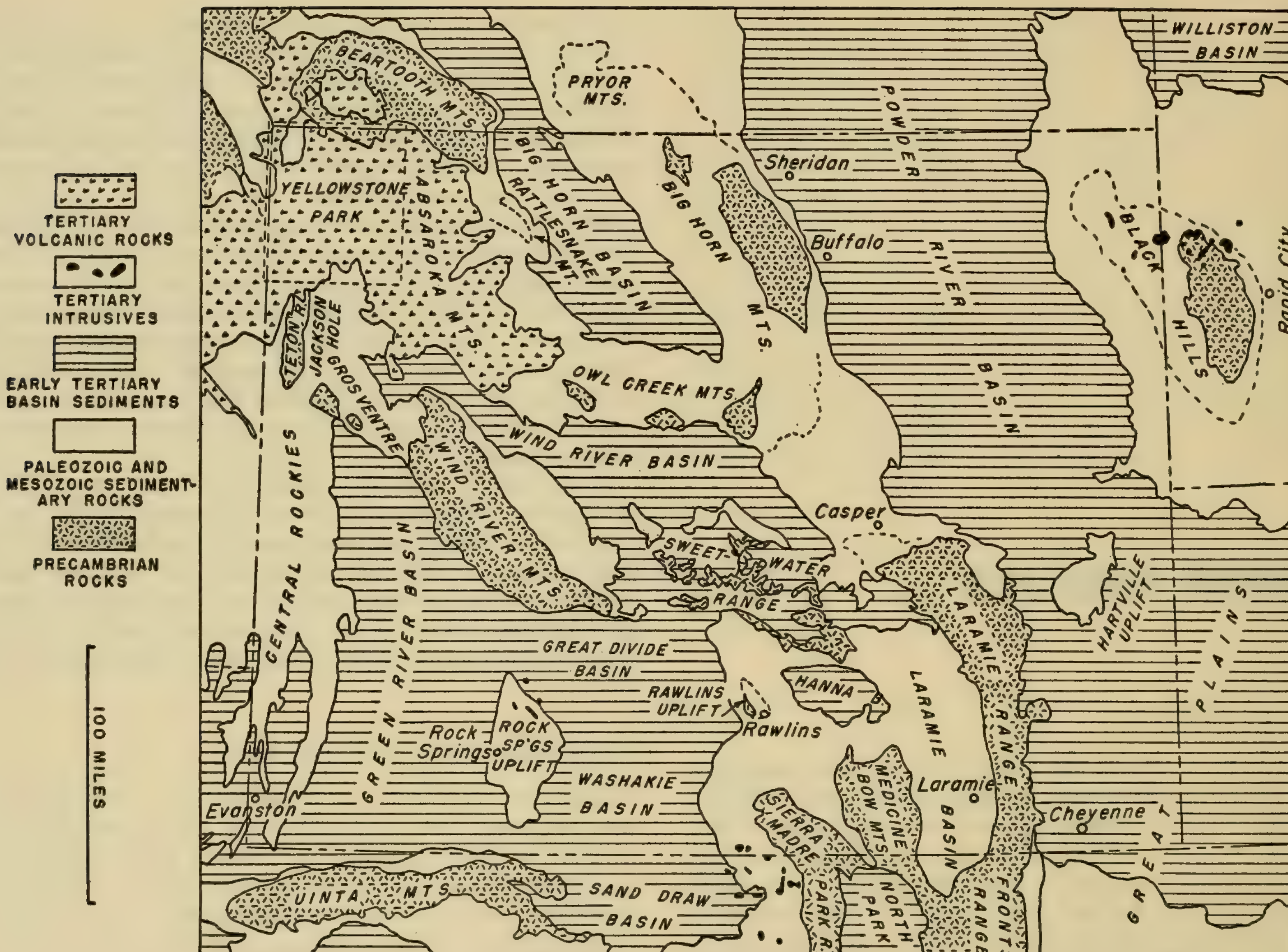


Fig. 24.1. Index map of Wyoming. Certain small Tertiary basins not shown.

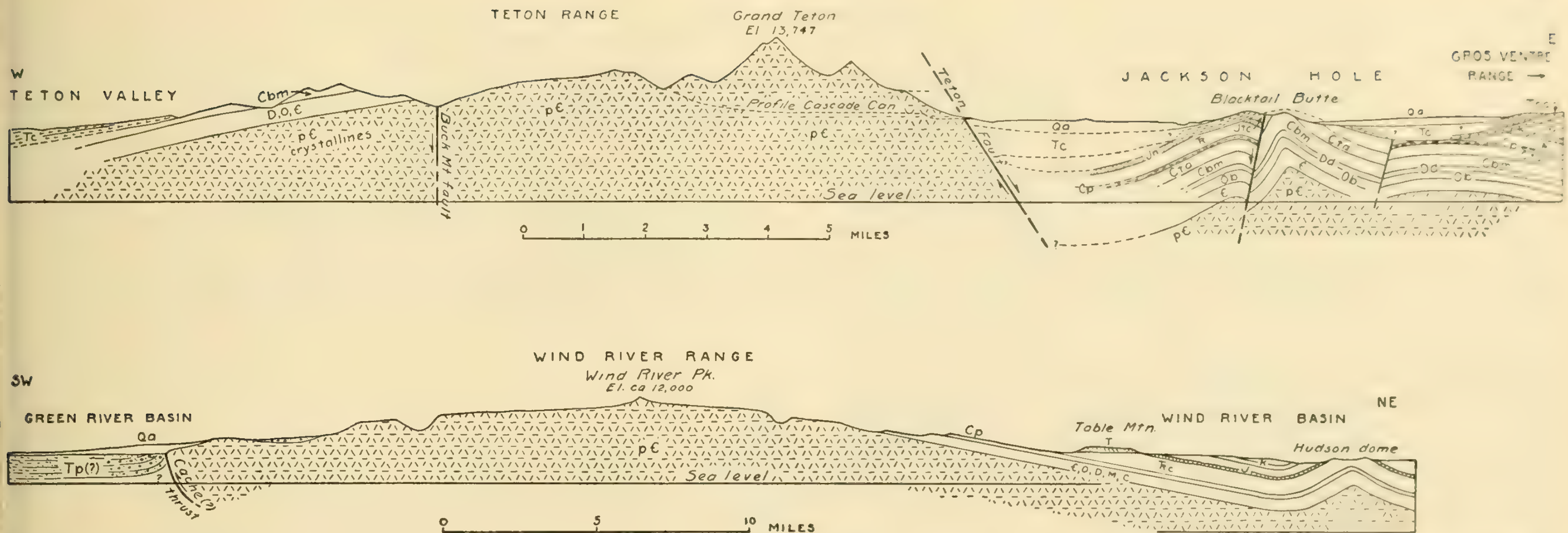


Fig. 24.2. Cross sections of the Teton and Wind River Ranges. The west slope of the Tetons is after Horberg (1938), the Blacktail Butte and Gros Ventre geology after Foster (1946), and the eastern slope of the Wind Rivers is after Branson and Branson (1941). Other parts are by the author. Ob, Big Horn dolomite; Dd, Darby formation; Cbm, Brazer and Madison limestones; Cta, Tensleep and Amsden; Cp, Phosphoria formation; Ꞧc, Chugwater formation; Jn, Nugget sandstone; Jtc, Gypsum Spring and Twin Creek formations; Tp, Pass Peak (middle Eocene); Tc, Camp Davis (uppermost Miocene); Tca, andesites of Camp Davis formation.

Hills. Along their southwestern flank, however, steep upturning and overthrusting is the rule. The Wind River Range is separated from the Gros Ventre by a broad sag or saddle in which most of the Paleozoic and Mesozoic formations are preserved and in which folds and faults of considerable magnitude occur (Richmond, 1945). The Gros Ventre Range is separated from the Tetons by a broad and picturesque valley, Jackson Hole, which trends north and south. The depression is due mainly to late Cenozoic block faulting, and the Laramide structural setting between the two ranges is not known. The post-Laramide faulting has been discussed in Chapter 22, and will be mentioned again in Chapter 30.

A cross section into the Gros Ventre Range from the facing Hoback Range has already been presented (Fig. 22.9), and the structural relations of the two ranges discussed. Other sections of the Tetons, Gros Ventre, and Wind River ranges are given in Fig. 24.2, which by inspection should explain the broad features of each.

The southwest flank of the Wind River uplift has been traversed seismically by Berg and Wasson (1960), and they report a thrust that dips as low as 18 degrees and carries under the range about 8 miles. The amount of vertical uplift in the Wind River Mountains is in excess of 35,000 feet.

BEARTOOTH RANGE

The Beartooth Range extends from southern Montana into northern Wyoming. Its northeast front is uplifted and generally overthrust northeastward, whereas the southwest front of the Wind River Range is apparently overthrust southwestward. A number of porphyry intrusions are

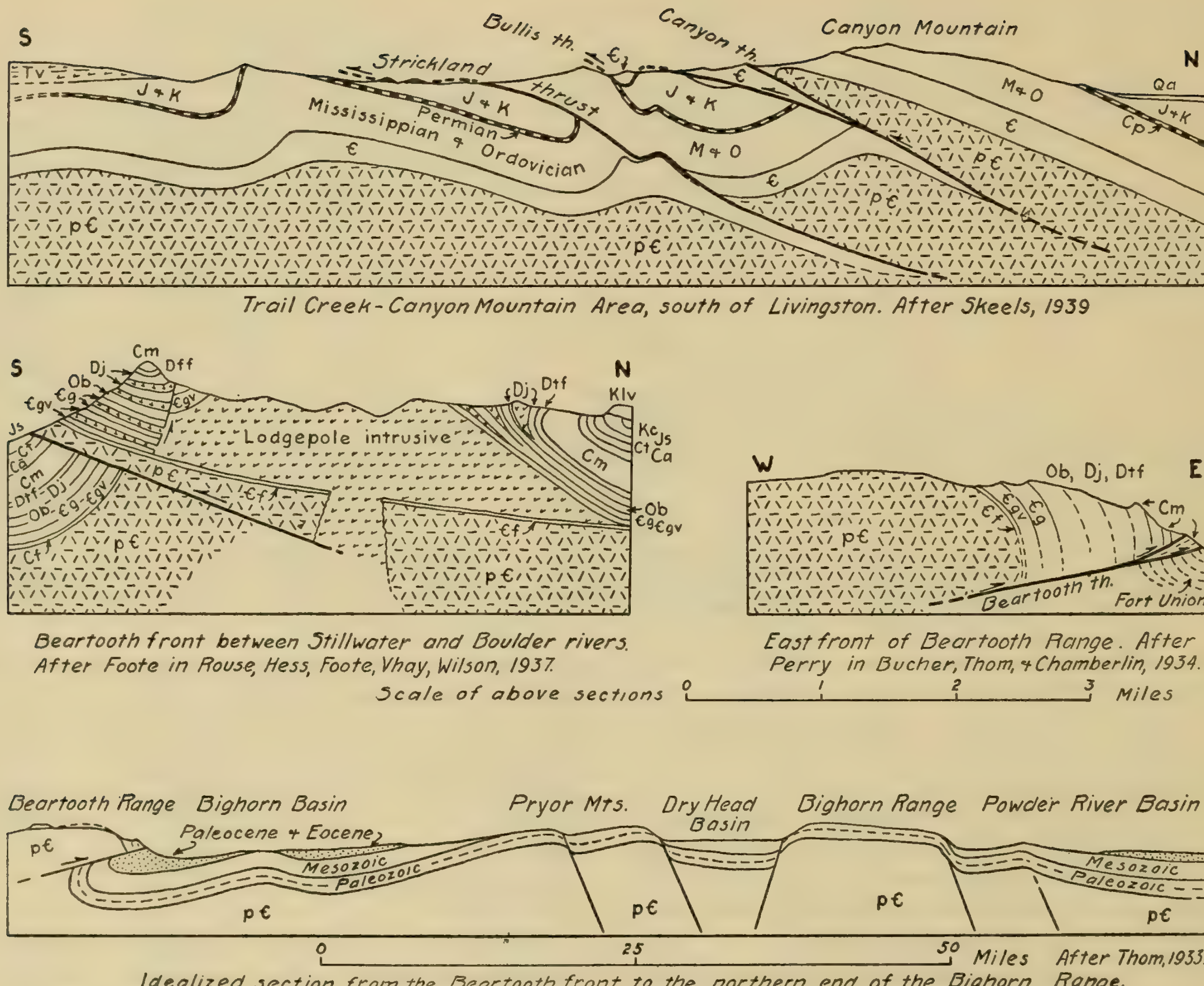


Fig. 24.3. Cross sections of the front of the Beartooth Range and adjacent basins.

in close proximity to the Beartooth thrust. They were intruded before the thrusting took place and have been cut and displaced by the fault or tears associated with it. See Lodgepole intrusive, Fig. 24.3. The intrusions are in the form of small sills principally in the Cambrian strata, nearly horizontal sheetlike masses not far below the Cambrian strata in the Precambrian and laccoliths. The latter are found near the mountain front where the Nye–Bowler lineament is closest.

The northwest end of the Beartooth Range and hills in the vicinity of Livingston, Montana, are structurally complex. The northward flowing Yellowstone River bounds the range on the west, but extending northwestward beyond are low mountains that link with the Bridger Range. The northeast front of the Beartooth Range is generally bounded by a low-angle thrust dipping into the range, and the thrust sheet has moved northeastward. In the Livingston area, however, several thrust sheets from in front of the main mountain block have moved southward and have been resisted by a corner of the "North Snowy block" (Lammers, 1937). See upper cross section, Fig. 24.3. The thrusting may have been preceded by a stage of folding and erosion which could correspond with the post-Lance and pre-Fort Union unconformity (Skeels, 1939). The thrusting itself may correspond to the post-Fort Union and pre-Wasatch unconformity in the Livingston basin. See discussion of the Beartooth thrust in Chapter 23.

Foose (1960) has treated the Beartooth Range as a rectangular block primarily elevated above adjacent basins and secondarily affected in places by horizontal transport of its marginal rock masses. At the northeast (Bear Lodge) corner the vertical structural relief is 15,000–20,000 feet, and in the absence of confinement, he concludes that the mountain mass has moved outward on the adjacent basin as much as 10,000 feet. The movement was facilitated by such secondary structures as bent high-angle faults, tear faults, and imbricate thrusts.

OWL CREEK AND WASHAKIE MOUNTAINS

Rattlesnake Mountain west of Cody and other smaller topographic features continue the Beartooth uplift southward, but on the west great accumulations of volcanics compose the mountain mass and extend southward for about 50 miles, where the Owl Creek Mountains appear. The volcanics spread northwestward, over considerable areas of sedimentary rock, and lay up on the southwest flank of the Beartooths. They form the Absaroka Range (lower section in Fig. 24.4). From under the volcanics a large asymmetrical anticline, the Owl Creek Mountains, appears, which extends generally eastward, and in places at least, is overthrust southward. See upper section in Fig. 24.4. The large anticline is broken by many faults and rendered further complex by small folds (Fanshaw, 1939). The shelf facies of Paleozoic, Triassic, and Jurassic rocks is essentially the same here as in the Big Horn and Wind River ranges. It is probable that the structures of the Owl Creek Mountains extend northwestward under the Absaroka volcanics so as to lie west of Rattlesnake Mountain and the Beartooth plateau, but the volcanics cover most of the area and little is known of the underlying rocks or structure.

Wise (1961) recognizes a primary vertical uplift of about 20,000 feet of the Owl Creek block, then gravity sliding of Mississippian strata away from the crest of uplift, where kegstone-like graben exist. The sliding toward the Wind River basin was unrestrained, and the lower extremities of the thrust sheets are much brecciated and grade into conglomerate lenses of the Eocene Wind River formation.

South of the Owl Creek Mountains and between it and the Wind River Range is the Wind River basin, which contains an instructive sequence of orogenic sediments. They are tabulated in Fig. 24.5. At the west end of the Owl Creeks and at the south end of the Absarokas is the Washakie Range, which has been studied in considerable detail by Love (1939). His account is representative of the Laramide history of the Wind River basin and adjoining ranges and is abstracted with minor changes as follows:

> The Owl Creek Mountains and the Washakie Range were folded and probably faulted at the close of Lance time and before the beginning of Fort Union deposition in the area to the northeast; the granitic core of the Washakie Range was exposed and being eroded when the upper part of the Fort Union formation was being deposited in the southwestern portion of the Big Horn basin; at the close of Fort Union time there was additional folding and probably faulting along the margins of the Owl Creek and Washakie ranges; the Pinyon conglomerate was deposited in the northwestern part of the Wind River basin dur-

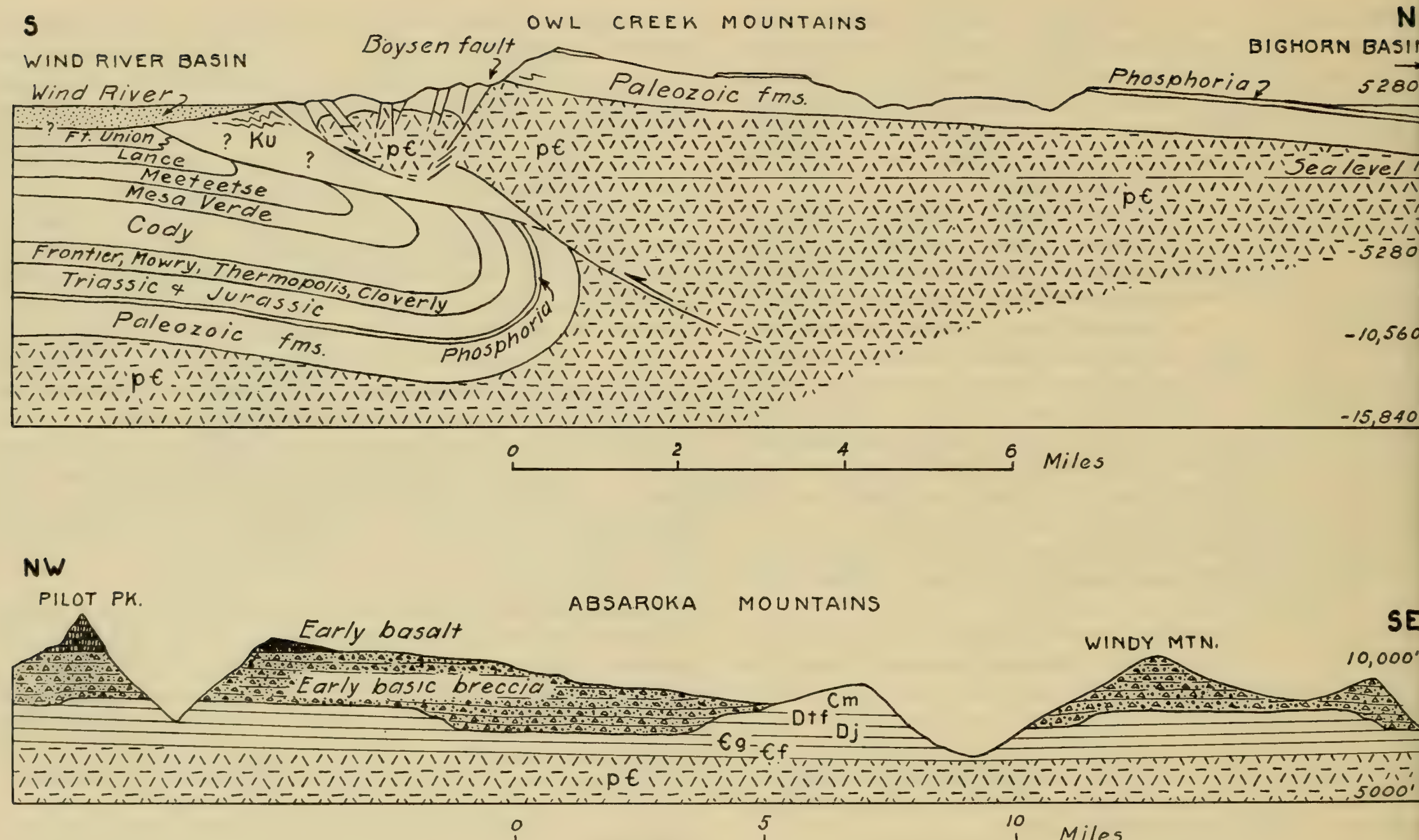


Fig. 24.4. Upper section through the Owl Creek mountains near the Wind River Canyon. After Fanshawe, 1939. Lower section along the southwest side of Clark Fork Valley in the Absaroka Mountains. After Rouse, 1937. pЄ, Precambrian crystallines; Єf, Flathead quartzite; Єg, Gallatin shale; Dj, Jefferson and Big Horn; Dtf, Three Forks; Cm, Madison; Ku, Upper Cretaceous undifferentiated.

ing Fort Union time; and a well-defined syncline north of the Washakie Range was drained by streams flowing from the west across the present site of the Absaroka Range.

The second pulsation of the Laramide orogeny came at the close of Fort Union time. The intensity and extent of folding and faulting are not known. During this and the preceding movement, the major structures of the Washakie Range developed.

Then followed the deposition of 1000 feet or more of early lower Eocene rocks (Indian Meadows) on a surface of high relief. The third pulsation of the Laramide orogeny is believed to have occurred at this time, and the klippen south of Coulee Mesa may be remnants of a thrust sheet pushed southward into the basin.

The fourth pulsation of the Laramide orogeny caused gentle folding along the northeastern flank of the Wind River Mountains. The early lower Eocene strata were eroded in places, and a broad southeastward-trending valley was formed between the Wind River and Washakie ranges.

Following this cycle of erosion, 500 feet of late lower Eocene rocks (Wind River) were deposited in this valley.

The fifth pulsation of the Laramide orogeny caused folding and thrust faulting along the center of the syncline between the Washakie and Wind River ranges. This was followed by the deposition of 1000 feet of middle Eocene rocks (Aycross) and the beginning of active Cenozoic volcanism in the general Absarokan region. Acidic and andesitic volcanic and pyroclastic rocks dominate.

The sixth pulsation of the Laramide orogeny resulted in gentle localized folding and some erosion after the close of middle Eocene time.

Deposition of 3000 feet of Oligocene (?) pyroclastic rocks (Wiggins), intrusion of plugs, extrusion of flows, and climax of Cenozoic volcanism. Acidic andesites dominate. Washakie and Owl Creek ranges were completely buried; Wind River and Bighorn basins were filled; Wind River and Bighorn ranges were partially buried.

The eighth pulsation of the Laramide orogeny caused folding in localized areas, recurrent uplift along parts of the buried Washakie Range, and erosion. Intrusion of dacite plugs and extrusion of flows followed.

The cross section of Fig. 24.6 shows some of the above relations.

HEART MOUNTAIN AND RELATED FEATURES

South of the Precambrian mass of the Beartooth Range and lying along the east front of the Absaroka Mountains are a number of relatively smaller features made up dominantly of Paleozoic limestones and dolomites. See Fig. 24.7. The anticlines known as Pat O'Hara and Rattlesnake

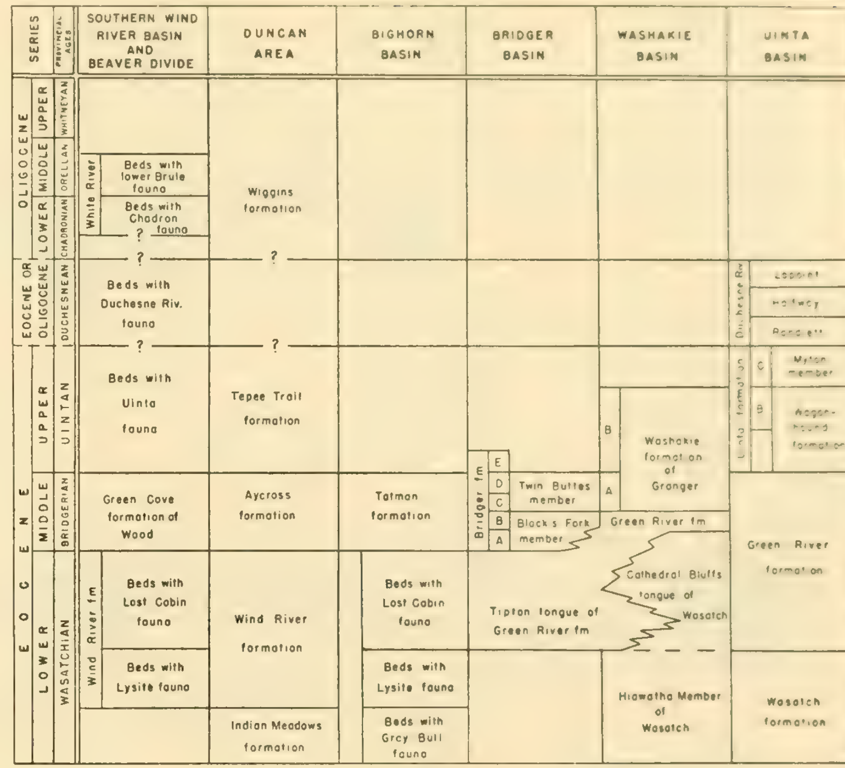


Fig. 24.5. Correlation chart of the lower Tertiary formations of south-central Wyoming and the Uinta basin in Utah. After Tourtelot and Nace, 1946.

Mountains, and Logan Mountain and Sheep Mountain are prominent. Two remnants of Paleozoic strata, Heart Mountain and McCulloch Peaks, consisting of large, irregularly disposed blocks rest on the Eocene Willwood formation. Heart Mountain is in the Big Horn basin at least 12 miles east of any possible root area, and McCulloch Peaks is over 28 miles. How these relatively small masses got where they are has proved a real mystery, and considerable has been written about them. Pierce (1941 and 1957) summarized the previous views and presented his own interpre-

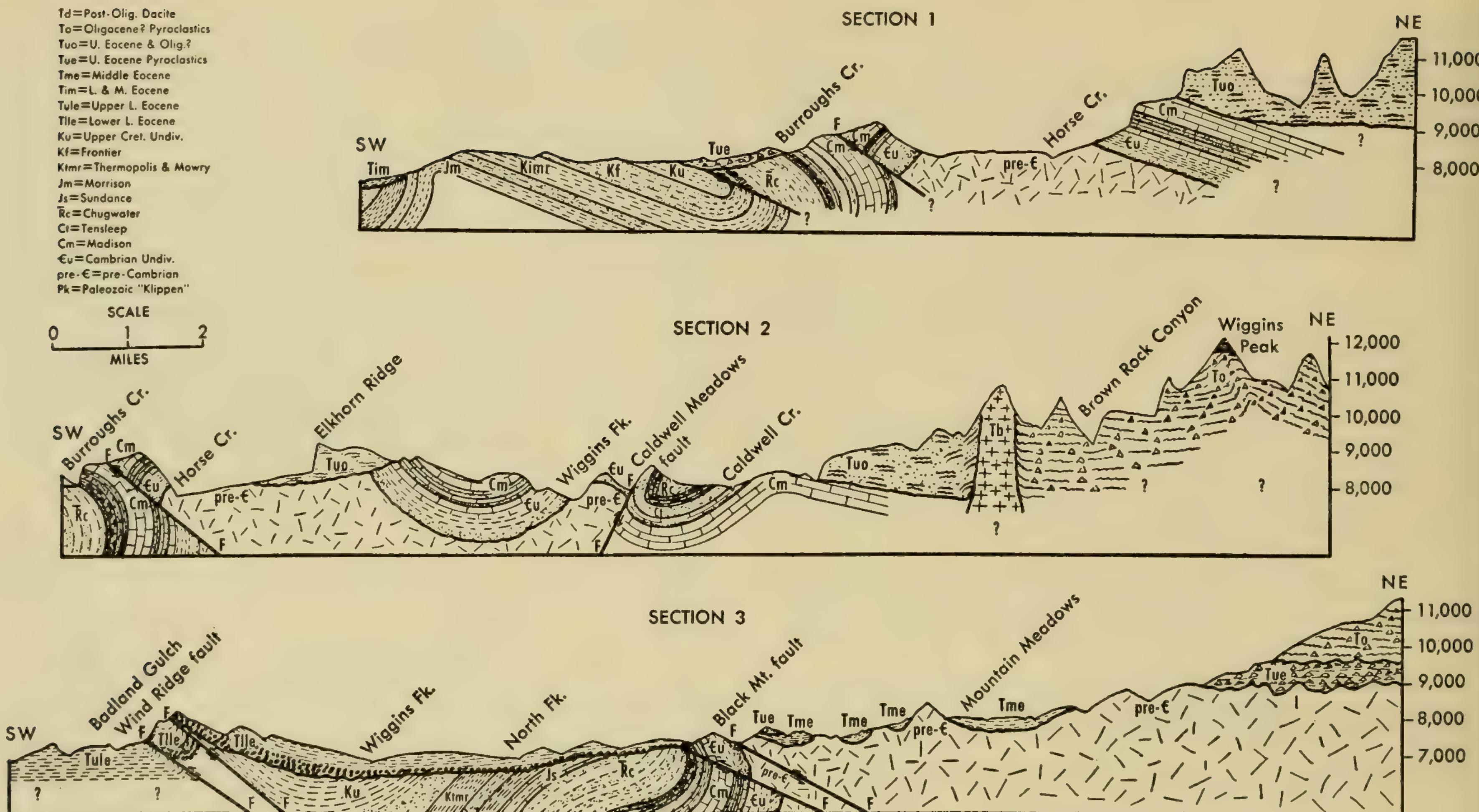


Fig. 24.6. Cross sections of the Washakie Range at the south end of the Absaroka Mountains. Reproduced from Love, 1939.

tation to the effect that the Paleozoic remnants of Heart Mountain, McCulloch Peaks, Logan Mountain, and Sheep Mountain are detached, gravity slide blocks as represented on Fig 24.7. He recognizes a second and immediately older thrust, the South Fork (mapped also as detached slide remnants) and the Logan and Sheep Mountain remnants to have slid on top of the South Fork remnants.

In broad outline the Heart Mountain fault of Wyoming is a nearly horizontal thrust whose overriding sheet was derived from a source without any known

roots, and whose frontal part has ridden across a former land surface. The suggestion is here made that this thrust and the near-by South Fork thrust are detachment thrusts or decollements, that is, they are sheets of sedimentary rocks which have broken loose along a basal shearing plane, have moved long distances probably by gravitational gliding, and have been deformed independently from the rocks below the fault plane.

The present remnants of the Heart Mountain thrust sheet include more than 50 separate blocks which range in size from a few hundred feet to 5 miles across and which are scattered over a triangular area 30 miles wide and 60 miles long. The rock formations represented in the thrust blocks comprise a very limited stratigraphic range, none being older than the Bighorn dolomite (Ordovician) and none younger than the Madison limestone (Mississippian). The maximum stratigraphic thickness of the formations involved is 1,800 feet, but these include the most competent group of beds in the sedimentary sequence in this area.

In the northwestern part of its known extent the Heart Mountain thrust plane follows the bedding of the rocks and lies at the base of the massive and resistant Bighorn dolomite and above the underlying Grove Creek formation (a thin unit at the top of the Cambrian sequence). Near the center of the area here described this bedding thrust plane changes abruptly to a shear plane that cuts stratigraphically upward across the Bighorn and younger formations; the thrust plane then passes southeastward onto and across a former land surface. The present thrust remnants on this surface are separated blocks that rest on rocks ranging in age from Paleozoic to Tertiary. See Fig. 24.8.

In the area of the bedding thrust the displaced sheet was broken into numerous blocks which became detached from one another by movement, with large spaces or gaps separating them. Thus by tectonic denudation the thrust plane was exposed at the surface. Associated with the events accompanying the thrusting was the rapid formation of a stream channel deposit, here named the Crandall conglomerate. Next there followed the deposition of the "early basic breccia." This blanket of volcanic rock, which is now in the process of being eroded, has preserved much of the geologic record pertaining to the development of the Heart Mountain thrust since middle Eocene time.

Pierce (1960) has more recently recognized the break-away point of the detached slide blocks.

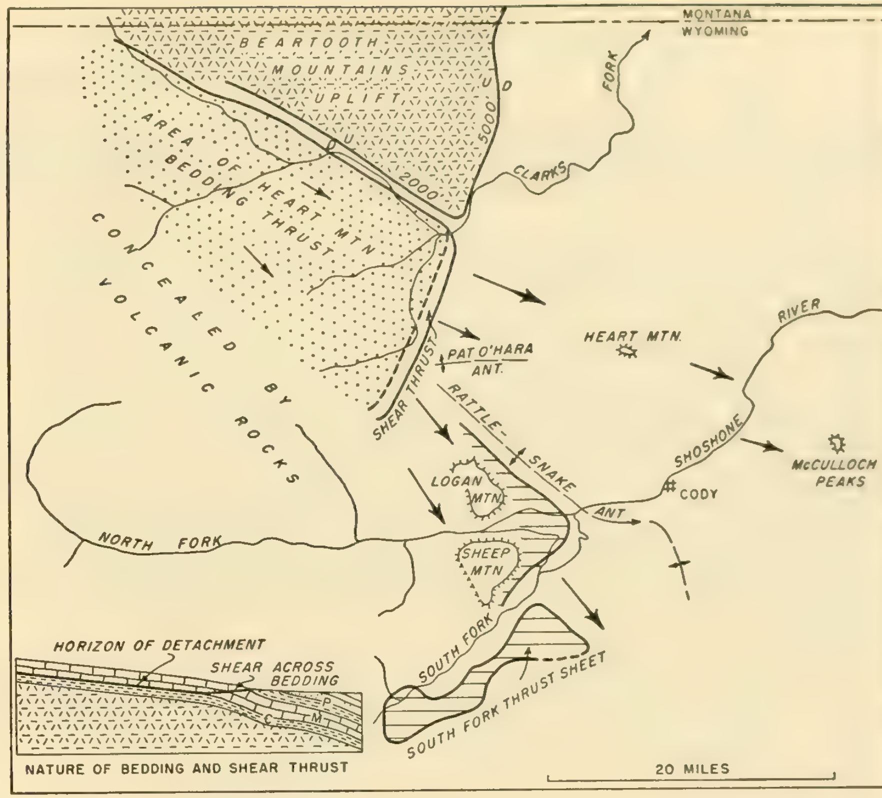


Fig. 24.7. Origin of Heart Mountain thrust, after Pierce, 1957. Dotted area is postulated area of bedding plane gliding of extensive sheet; zone marked shear thrust is where glide sheet cut across beds, and the area to east is where detached blocks glided 20 and 40 miles to form Heart Mountain and McCulloch peaks. South Fork thrust is older than Heart Mountain thrust, and Logan Mountain and Sheep Mountain are detached remnants of Heart Mountain thrust resting on South Fork sheet.

ABSAROKA RANGE AND YELLOWSTONE PARK

Breccia Series of the Absaroka Range

The Absaroka Range and Yellowstone Park comprise a large volcanic area which is made up of pyroclastic rocks and lavas. Two groups have been distinguished, each of which is composed of a lower acid breccia, a middle basic breccia, and an upper series of basalt sheets. Altogether, they are known as the breccia series. See columnar sections in the chart of Fig. 24.9. The early acid breccia was probably erupted just before the Heart Mountain thrust occurred, and the succeeding breccias and flows accumulated on a rugged surface, the local relief of which ranged from

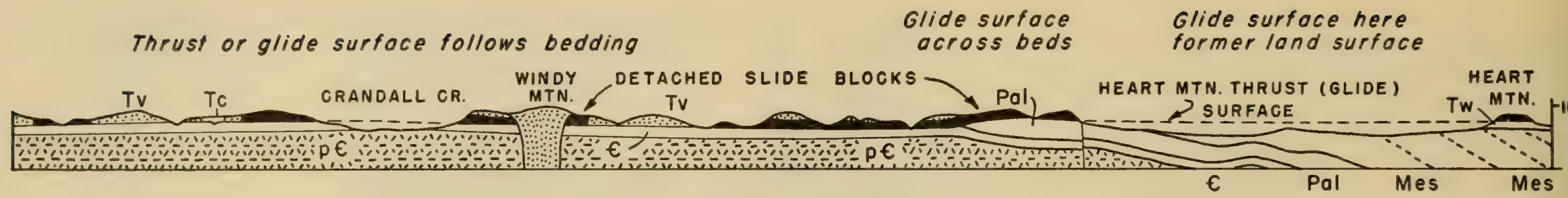


Fig. 24.8. Generalized section along Clarks Fork to Heart Mountain showing nature of thrust (glide) surface and detached glide blocks. After Pierce, 1957. Tv, mostly early basic breccia (middle Eocene); Tc, Crandall conglomerate (early (?) Eocene); Tw, Willwood fm. (early Eocene); Mes, Mesozoic and Paleocene strata; Pal, Permian to Ordovician strata. The Heart Mountain thrust blocks are from the Madison, Threeforks, Jefferson, and Big Horn formation. Vertical scale in thousands of feet.

1000 to 4000 feet. The maximum thickness of the series in any continuous section is 6500 feet (Rouse, 1937).

The pyroclastic rocks were erupted through hundreds of small vents and from a few volcanoes of moderate size. The basalt sheets are all fissure eruptions.

The breccia series have been divided into formations better suited for mapping purposes by Hay (1954) and Wilson (1959), and the succession of formations thus established is given under the column, Wood River, in Fig. 24.9.

		BIG HORN BASIN	ABSAROKAS (OLD TERMINOLOGY)	ABSAROKAS (WOOD RIVER)	WASHAKIE MTS.
MIOCENE				Intrusives: Rhyolite, Granodiorite, Andesite, Dacite, Rhyodacite	
OLIGOCENE			Late basalt flows Late basic breccia Late acid breccia	Wiggins fm. (volcanics)	
EOCENE	UPPER		?	?	Tepee Trail fm.
EOCENE	MIDDLE	Tatman fm.	Early basalt flows Early basic breccia	Pitchfork fm. (80% andesite)	Aycross fm.
EOCENE	LOWER	* Willwood fm.	Early acid breccia	Willwood fm.	Wind River fm. Indian Meadows
PALEOCENE		Polecat Bench fm.			

* Crandall conglomerate of Heart Mountain region is Lower Eocene.

Fig. 24.9. Tertiary formation of Big Horn basin and Absaroka Mountains.

Plutons of the Absaroka Range

In the northern Absaroka Range stocks, laccoliths, plugs, cone sheets, and radial dike systems occur and are closely related to the volcanic centers. The magma of the radial dikes moved horizontally outward. The rocks in general show a normal differentiation series from Olivine gabbro and basalt through diorite and andesite to sodic syenite and trachyte (Parsons, 1939).

A number of intermediate felsic stocks are known in the central area of the southern Absarokas, and these occur in striking alignment. The zone may have served first for the breaking through of the volcanic conduits and then later for the stocks that cut the breccias (Rouse, 1937).

The kinds of post-Wiggins intrusive rocks which Wilson (1959) mapped in the Wood River area are listed in Fig. 24.9.

Breccia Series of Yellowstone Park

The general region of Yellowstone Park was a basin in the time of accumulation of the first volcanics, the early acid breccias. Although partly surrounded by higher topographic features, it was a rugged surface like that of the early Absaroka Range but somewhat lower. It is estimated that in places about 1000 feet of the early acid breccias accumulated (Howard, 1937). Then the voluminous early basic breccias were erupted. These include breccias, agglomerates, tuffs, and flows of a more basic character with basalt predominating. They reach a maximum thickness of 4000 feet and have a wide distribution from the Absaroka Range through northern Yellowstone Park to the Gallatin Range in Montana

northwest of the park. The Washburn Range within the park is formed entirely of the early basic breccias.

A trachyte was possibly extruded next, approximately along the course of the Yellowstone River. Then over the early basic breccias, but nowhere over the trachyte, were poured out a great series of basalt flows 1200 feet thick. These basalts form many of the higher flat-top summits in the northern part of the Absaroka Range, but are overlain by later deposits to the south. They are distributed widely in eastern Yellowstone Park, and they are an important horizon marker because they separate the early breccias from the later.

Renewed explosive activity resulted in the accumulation of the late acid breccias previously mentioned. They are limited chiefly to those portions of the Absaroka Range that lie within the Park and extend westward to Yellowstone Lake.

A period of erosion evidently followed, and the late basic breccias 2500 feet thick were deposited over an irregular surface. They form extensive plateau areas in southeastern Yellowstone Park and make up chiefly the southern half of the Absaroka Range but are exposed sparingly over the late acid breccias. Where the late acid breccias are absent, the late basic breccias rest directly on the early basalt sheets. The eruption of more basalt flows closed the period of late basic breccia volcanic activity.

The last of the breccia series, which includes the early and late breccias and the basalts, was an andesite outpouring in the southeastern part of the park. It is now preserved in the higher peaks there.

The following is Howard's summary (1937) of the post-breccia history of Yellowstone Park.

Post-Breccia Faulting. Study of the Washburn Range indicates that the next event of major importance was extensive faulting of the great series of volcanic rocks previously described. It appears to have been this faulting that gave the Washburn Range a relief so great that it could not be buried by the later rhyolite floods. Presumably, other inequalities of the old basin floor are attributable to faulting at this period, but only where the later rhyolite failed to bury the inequalities, or where post-rhyolite erosion has later uncovered them, can the evidence of faulting be studied.

Post-Breccia Erosion. The faulting of the great masses of pre-rhyolite volcanic formations was associated with a long period of erosion, sufficiently important to deserve special mention. Locally, at least, a gently rolling topography was developed on the breccias within the park. Thus, the erosion contact between the breccias and the overlying rhyolite, where exposed for a distance of 8 miles in the walls of Yellowstone Canyon, from Broad Creek almost to Tower Creek, is gently undulating. Elsewhere, the relief is much greater, but how much of it is due to faulting is unknown. Presumably, the faulting took place progressively over a considerable period, and erosion must have accompanied the movements. Whether sufficient erosion preceded the faulting to produce a faint relief, which was then locally intensified by uplift, or whether strong relief due to early faulting was not effaced by the erosion that elsewhere produced a gently rolling topography, is not clear from the evidence obtained.

After the faulting and erosion, the Yellowstone basin, its dissected sides and floor now composed partly of pre-Tertiary rocks of all kinds and, partly, of Tertiary volcanic rocks, received the floods of late Tertiary rhyolite lavas. The rhyolite floods were locally preceded by basaltic extrusions.

Early Canyon Basalts. Basalts are found locally under the rhyolite, below the level of the surrounding breccias, and indicate a period of eruption later than that represented by the basalt that closed the period of breccia accumulation. The early Canyon Basalts were probably poured out after the faulting of the breccias and after the extended erosion period associated with that faulting. They are exposed in patches along Yellowstone Canyon and in the canyon of Gardiner River, in the northwestern part of the park.

Rhyolite Floods. There now occurred one of the most remarkable events in the history of Yellowstone Park, for enormous floods of rhyolite lava filled the lowlands of the earlier landscape to depths of a thousand feet or more, swept around the Washburn Range and other highland areas, which stood as islands in the lava sea, lapped against the foothills of the encircling ranges, and continued an unknown distance to the west, where the mountain rim is lacking. Today, the lava plateau terminates a short distance outside the Park in a steep scarp of uncertain origin, which drops sharply to the lower Snake River Plains.

Certain basalts in the northern part of the Park, at the edge of the basin, may have been extruded during pauses in the extrusion of the rhyolite.

Post-Rhyolite Faulting. Following the extrusion of the rhyolite, the broad, level plateau surface was broken by block-faulting, perhaps a result of settling in response to the withdrawal of the vast quantities of magma from below. Many of the lake and hot-spring basins, and many of the "topographic fault blocks" visible on the contour maps, may have been formed at this time.

Post-Rhyolite Erosion. There next ensued a period of erosion, the extent of which remains an unsolved problem. The crispness of most of the block units in the topography, however, suggests slight denudation of the park area as a whole, but a few deep valleys, such as the Lamar Valley in the north, may have been eroded. The carving of the Grand Canyon of the Yellowstone River may have begun at this time or during the first half of the Pleistocene. Its present depth, however, was attained during the late Pleistocene. The scarp that limits the rhyolite plateau to the west was presumably fashioned at this time, for its

base is submerged by basalts, which are probably the equivalents of the Late Canyon Basalts.

Late Canyon Basalts. Erosion was then followed by another period of basalt extrusion, these basalts being the most recent flows of the Park. They are found largely in the northern part of the Park, in Lamar, Yellowstone, and Gardiner valleys, and at a few places on the broad interstream uplands. Other patches are preserved on the uplands in west-central Yellowstone. The basalts of the Snake River Plains, which crowd against the western scarp of the rhyolite plateau, are probably of the same general age.

BIG HORN RANGE AND BIG HORN BASIN

Divisions of Big Horn Range

Overall the Big Horn Range is a great anticlinal fold, steep to overturned to overthrust on the east, and gently dipping on the west. Examine Fig. 24.1. The range is arcuate in plan view and terminates in the Pryor Mountains on the northwest and the Owl Creek Range on the southwest. The Precambrian crystalline rocks on which the Paleozoic strata rest are exposed in three areas in the core of the range, and serve as natural divisions.

At three points near the center of curvature in the central division along the east front of the range, blocks of the range, including the crystallines, have been thrust out upon the Cretaceous strata. The main block is clearly bounded by tear faults.

On cross sections published by Hoppin (1961), overturning, thrusting, and a detached slide mass are shown. It appears evident that sharp uplift and upturning of the beds are the primary movements and then that secondary gravity movements have resulted in downhill overturning of the beds, tear faults, and small-scale thrusting.

As the axis of the great fold in the central division is traced northward, it plunges, and the dips on the northeast flank become gentle. Beyond, in the northern division, the asymmetry is reversed, and the crystalline rocks are exposed close to the southwestern flank. Here steep dips, overturning, and even thrusting to the southwest occur. The northern division is further distinguished by sharp flexures which trend northwestward, northward, and eastward.

In the southern section, the trend of the mountain axis curves from a north-south direction to a southwesterly one; but in spite of this change, the smaller structures within the range maintain the northwesterly directions that dominate the northern division. The marginal folds and faults trend dominantly to the northwest, and the dips are steeper on the southwest sides of these small folds (Bucher, 1934).

The Tensleep fault cuts across the Big Horn Range from the town of Tensleep on the west to the Horn on the east, and separates the central from the southern divisions. As the range was uplifted and the central division developed asymmetrically eastward and the southern asymmetrically westward, the Tensleep fault came into existence (Wilson, 1938; Demorest, 1941). Relations are complex along the fault, but they point to a downthrow on the south side.

Laramide History

The Laramide history of the Big Horn Range has been summarized by R. P. Sharp (personal communication) for the writer. According to him, a series of coarse to bouldery fans (the Kingsbury conglomerate) composed primarily of Precambrian debris, were built up along the east base of the central Big Horn Mountains in Early Tertiary time. This was the section of greatest uplift, and the fan debris was presumably coarser and thicker here than elsewhere. Subsequently, the Paleozoic beds of the mountain front were thrust eastward against and over the gravel, and erosion during the remainder of the Cenozoic has gradually etched out the thickest and coarsest parts of the fan deposits so that they form prominent ridges in the present landscape. At least three periods of Laramide uplift of the range are indicated: (a) An uplift which produced the Kingsbury conglomerate. Faulting probably occurred during this uplift. (b) A second uplift, also probably attended by faulting, which deformed the Kingsbury and produced the coarse granite-boulder gravel. This uplift may possibly have been accompanied or closely followed by alpine glaciation. (c) A third, postgravel, uplift marked by thrust faulting toward the east in the central segment of the range.

Pryor Mountains

Northwest of the Big Horns the Paleozoic strata rise once more by means of two pairs of flexures to form the Pryor Mountains. One pair trends

east-west, and the other north-northwest to divide the uplift into four units. In three of these units, the beds rise toward a high point near the northeast, beyond which they drop off abruptly. Most of the flexures have ruptured to produce faults of moderate displacement. Thom (1923) and later Blackstone (1940) have concluded that the faults curve under the uplifted blocks at depth and have resulted from horizontal compression. The pliable sedimentary veneer flexes first over the scarp of the crystalline rocks and later, when displacement becomes sufficient, it breaks to reveal the deep-seated fault. See Fig. 24.10.

Big Horn Basin

The Big Horn basin is underlain in its deepest parts by 2500 to 3200 feet of Paleozoic strata, by about 1500 feet of Triassic and Jurassic strata, by 7000 to 9000 feet of Cretaceous strata, and in the central and western parts by several thousand feet of Paleocene and Eocene strata. For a review of the formations, see *Wyoming Geological Society Seventh Annual Field Conference Guidebook*, 1952. As indicated in earlier parts of this book, the Wyoming region, including the Big Horn basin, was a shelf area of sedimentation until Cretaceous times, when considerable subsidence occurred adjacent on the east to the active Cordilleran geanticlinal area that extended through Utah and eastern Idaho. See the paleotectonic maps of the Early and Late Cretaceous.

With the elevation of the ranges surrounding the Big Horn Basin, its sediments were thrown into many folds, all trending in a northwest direction. The Early Tertiary strata probably obscure many folds in the central part of the basin, for the anticlines and synclines are known only in a broad marginal belt. Those on the east side have steep flanks facing the Big Horn Mountains. The major anticline on the west side, the Rattlesnake Mountain, is asymmetrical toward the west, but the smaller folds do not have any regular symmetry. Some have steeper flanks toward the basin, some are about symmetrical, and some are dome shaped. The anticlines and domes are nearly all oil or gas producing. Two of the anticlines, especially, are cut by numerous, small, high-angle faults in a transverse direction. These are the Elk Basin and Garland anticlines in the northern part of the basin. The deepest part of the Big Horn basin, according to

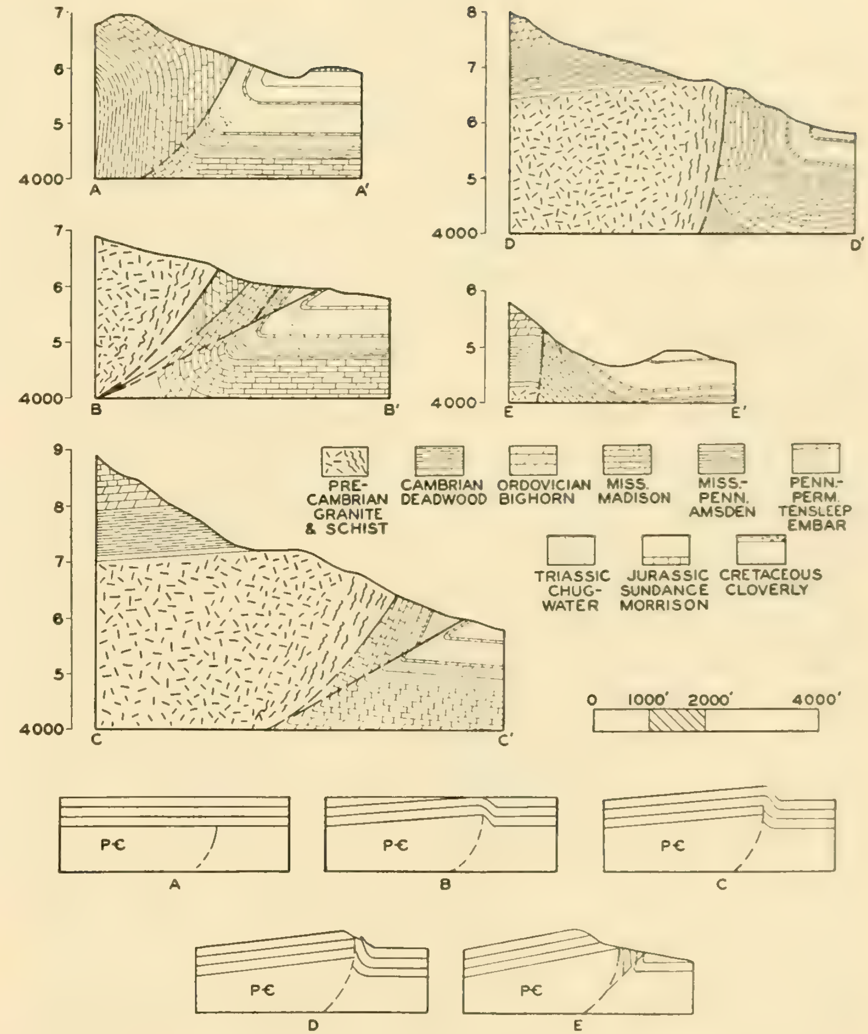


Fig. 24.10. Cross sections of the frontal faults of the tilted blocks of the Pryor Mountains. The lower diagrams, A to E, illustrate the theory of origin. Taken from Blackstone, 1940.

geophysical prospecting, is somewhat west of the geographical center.

The folds of the Big Horn Basin, according to Fanshawe (1947), are due to an interplay of two forces. The Precambrian basement was faulted as it adjusted to Laramide mountain building on both sides, and the Paleozoic and Mesozoic strata flexed over the fault scarps. Also, as the sides of the basins were upturned, the upper beds of the basin were crowded and buckles developed.

Map, Fig. 22.4, shows the Big Horn Basin to have come into existence in Montana time, and Van Houten (1952) notes that Precambrian rock had been exposed in places in the surrounding ranges by late Paleocene time (Fig. 22.5). Sandstone, mudstone, and coal beds accumulated to a thickness of 7000 feet just east of the Beartooth front during Paleocene time.

The early Eocene Willwood formation overlies older beds unconformably at the margin of the basin, and this time is taken as one important deformation of the surrounding uplifts. As previously noted, the detached blocks of the Heart Mountain thrust (?) rest on the Willwood. See Fig. 22.6. The Willwood is spread widely over the Big Horn basin.

Middle Eocene time saw the accumulation of the Tatman formation, which is almost free of volcanic debris except at Lysite Mountain at the south end of the basin. West of the southern half of the Big Horn basin the Tatman is overlain by more than 1000 feet of volcanic debris of the Absaroka Range. Remnants of the volcanics are noted elsewhere, and it is postulated by Van Houten (1952) and Love (1956a,b) that sedimentation continued after Tatman time.

By late Eocene time the Big Horn basin had sunk relative to the uplifts on either side about 17,000 feet. About 9000 feet of the depression had been filled. Yet all the while, Mackin (1947) and Van Houten (1952) contend, the climate had not been changed, and the orogenic debris accumulated in a warm, humid lowland near sea level. In middle Cenozoic time gradual *regional* uplift occurred. Pediments were widely cut in the uplands and the lowlands were broadly alluviated, producing an extensive graded surface. By late Cenozoic time further regional uplift and increased aridity initiated the present cycle of erosion, and the graded surface was widely dissected.

Intrusive Rocks

A belt of Laramide intrusions extends across the Black Hills about at the north end of the exposed Precambrian core. Most have been considered laccoliths or modified laccoliths, such as Ragged top laccolith (Fig. 24.11), Bear Butte, Deadman Mountain, Cook Mountain, Whitewood Peak, Black Buttes, and Devils Tower (Robinson, 1956). Within the Precambrian basement complex the intrusions are chiefly sills and dikes, and by charting the base of the Cambrian sandstone Noble *et al.* (1949) have shown that the sedimentary rocks overlying the Precambrian have been domed in two places notably, and believe that intrusive stocks are the cause. Some of the so-called laccoliths are believed to be stocks.

BLACK HILLS AND POWDER RIVER BASIN

General Characteristics of Black Hills

The Black Hills rise island-like several thousand feet above the surrounding Great Plains in western South Dakota and northeastern Wyoming. They are the easternmost of the outer ranges of the Rockies, and in point of Laramide structure involving the sedimentary rocks, perhaps the simplest. Their ridges, peaks, and valleys are the erosional remains of a broad dome, some 120 miles long and 60 miles wide. A Precambrian core of crystalline rocks trends nearly north-south and is flanked by upturned and truncated Paleozoic and Mesozoic strata. The broad anticline trends and pitches northwestward beyond the crystalline area. The east flank is fairly steep, with dips up to 45 degrees and more; the broad top is fairly flat; and the west flank is fairly gentle, with dips of a few to 20 degrees. Four geomorphic units are distinct, namely, the central Precambrian core of fairly rugged mountains, a plateau area in the west central part that is formed of Paleozoic limestones not yet stripped from the Precambrian rocks, a remarkably continuous strike valley around the Hills eroded in Jurassic and Triassic strata between the inner Paleozoic formations and the outer Cretaceous sandstones (Fig. 24.11), and a bold, inward-facing hogback held up by the Fall River and Lakota Cretaceous sandstones. The strike valley is known as the Red Valley from the red

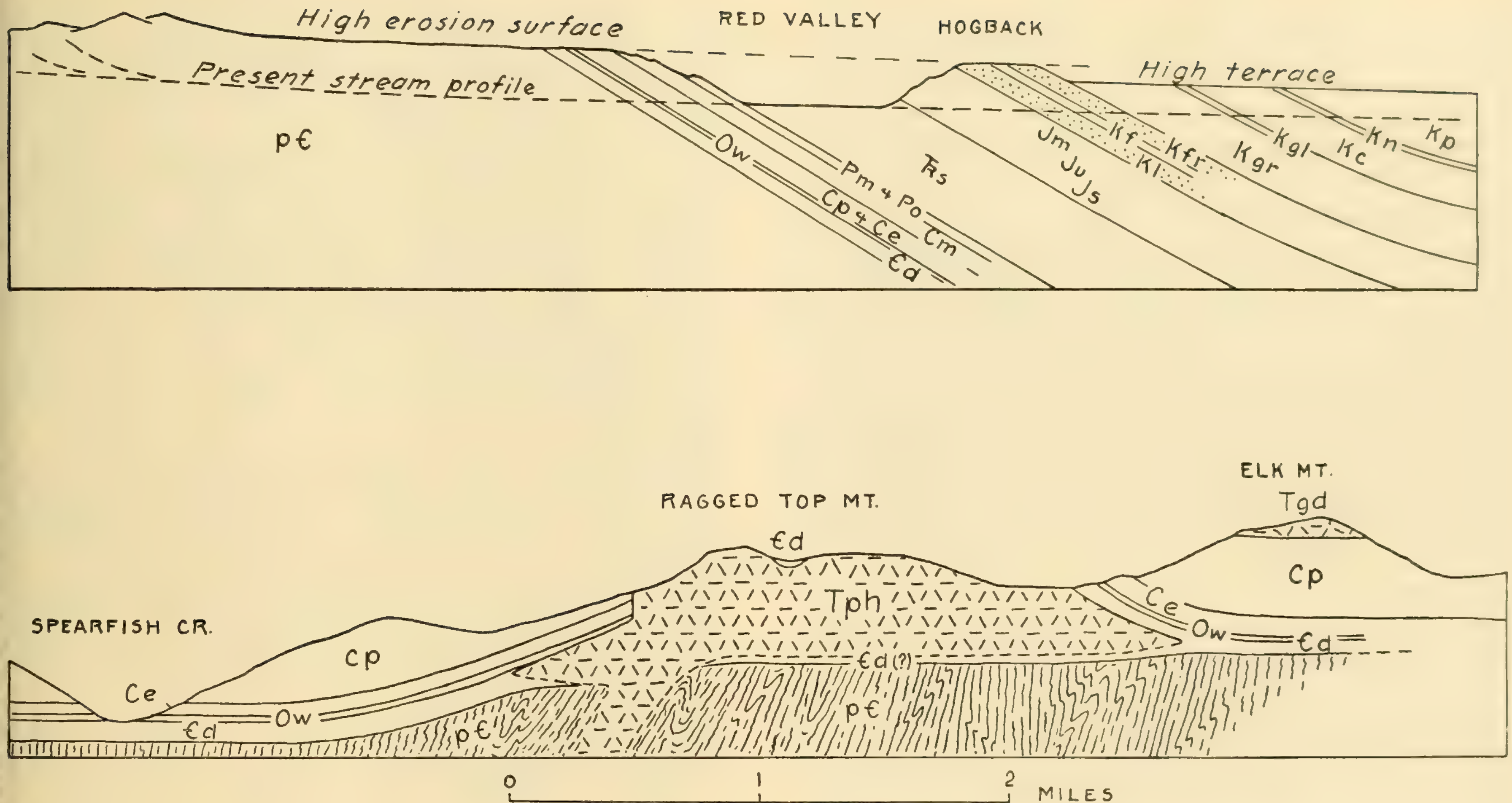


Fig. 24.11. Generalized cross section of the east front of the Black Hills just south of Rapid City, and cross section of the Ragged Top laccolith. Lower diagram after O'Harra, 1933. €d, Deadwood fm.; OW, Whitewood ls.; Ce, Mississippian Englewood ls.; Cp, Mississippian Pahasapa ls.; Cm, Pennsylvanian Minnelusa ss.; Pm and Po, Permian Opeche fm. and Minnekahta ls.; T̄Rs, Spearfish fm.; Js, Sundance fm.; Ju, Unkapapa ss.; Jm, Morrison fm.; Kl, Lakota ss.; Kf, Fuson sh.; Kfr, Fall River ss.; Kgr, Graneros sh.; Kgl, Greenhorn ls.; Kc, Carlile sh.; Kn, Niobrara sh.; Kp, Pierre sh.

Triassic Spearfish shales that principally underlie it, and also as the Race-track (Darton and Paige, 1925; O'Harra, 1933).

The Precambrian rocks consist of highly folded schists intricately invaded in the southern hills by large and small masses of granite. Laramide plutons intrude the Precambrian in the northern part, where the great Homestake gold mine is located, and, as some believe, are responsible for the ore deposits in large part.

General Characteristics of Powder River Basin

West of the Black Hills and between them and the Big Horn Range is the broad Powder River basin, floored by the Cretaceous, Paleocene, and Eocene beds. The Early Tertiary deposits are 10,000 feet thick in the deepest part of the basin, as indicated by seismic prospecting, and over most of the basin no reversals of dip, i.e., gentle anticlines or synclines, have been found. Only along the east flank of the Big Horns do any folds occur. Consult *U.S. Geological Survey Preliminary Map 33*. The very productive Salt Creek anticline is at the southern end of this belt. Darton estimates the strata were uplifted 9000 feet in the Black Hills, so the structural relief between the bottom of the Powder River basin and the top of the Black Hills is in the neighborhood of 20,000 feet.

Age of Uplift

The age of the uplift can be only approximated, because no Paleocene or Eocene overlaps exist. Those deposits of Laramide age that might have been in part derived from the Black Hills are now in surrounding areas fairly distant from the uplift and separated from it by a wide Cretaceous belt of outcrop. The doming could have started in latest Cretaceous time, with the deposition of the Fox Hills and Lance beds in the Powder River basin and around the north and northeast ends; and the distribution of the Fort Union and Wasatch beds, partly around the uplift and especially to the northeast, seem to indicate that the uplift had occurred and was furnishing some of the sediments that were accumulating.

Post-Laramide History

By early Oligocene time, erosion had trenched the uplift almost as deeply as now, and a mountain and valley surface of at least 1500 feet relief existed. Then the regimen of erosion changed to one of aggradation coincident with the change through central Wyoming and the Great Plains, and even in the early, deep valleys of the Black Hills, lower Oligocene beds began to accumulate (Darton and Paige, 1925). Deposition in these mountainous valleys lagged until middle Oligocene, whereas it was taking place in early Oligocene on the Great Plains (Fillman, 1929). The sediments may have reached such a thickness that all but the highest features of the range were buried, judging from the elevation of the White River beds to the east of the uplift, but if so they have since been removed within the hills except in small, protected patches. With the renewal of erosion, a drainage pattern, in details slightly at variance with the old, has failed to clean out all the Oligocene deposits, and has left them in places, forming low ridges and also extending down nearly to present valley bottoms. The surface upon which the middle Oligocene deposits accumulated in the hills has been called the Mountain Meadow (Fillman, 1929).

The Great Plains on the east of the Black Hills are covered by several formations ranging in age from lower Oligocene to Pliocene, and within this group are several disconformities. Some geologists have related the disconformities to uplifts in the Black Hills, but Mackin (1947) believes that the dominant form was a graded surface—erosional in the uplift and depositional on the peripheral regions. With regional uplift, in mid-Tertiary time, and associated change in climate to aridity, the graded surface was dissected to produce the landforms of today.

SWEETWATER RANGE

Extending westward from the north side of the Hanna basin to the southeast end of the Wind River Range is a series of hills, most of which are islands of Precambrian rock in Miocene and Oligocene sediments. Sufficient Paleozoic and Mesozoic strata are also exposed to indicate that the Precambrian islands demarcate the position of the core of a former great range, extending in general in an east-west direction through central Wyoming (Fig. 24.12). It probably was traversed obliquely by several sharp folds that trended in a northwest direction and which cast the bor-

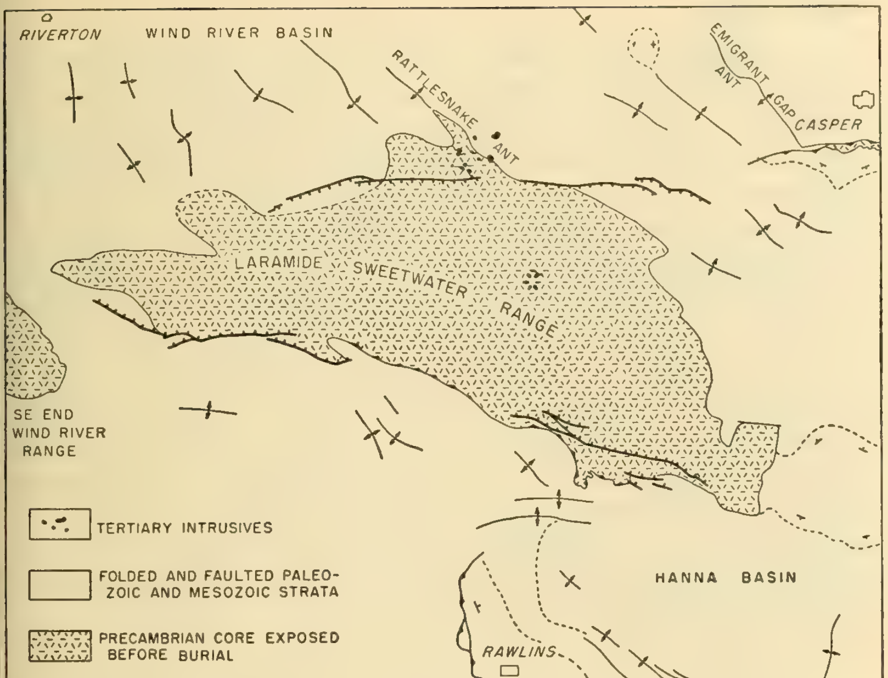


Fig. 24.12. Laramide Sweetwater Range and Late Tertiary normal faults. Somewhat after Blackstone, 1951. Thick hachured lines are the Late Tertiary faults. All others are Laramide. The range subsided and was covered by Mid-Eocene, Oligocene, and Miocene sediments and volcanics. Later erosion has exposed many peaks of the old range which are called the Granite Mountains. The Hanna and Wind River basins contain thick Early Tertiary deposits.

ders of the Precambrian core into a jagged pattern with a decided northwest fabric. It is clear that the range was elevated in the general Laramide revolution, and that the sedimentary veneer and probably much Precambrian rock was removed before the range started to sag. It was a singular phenomenon in Wyoming and Colorado, because all other Laramide ranges have remained as strong relief features until today, but similar to a Laramide uplift in southwestern Montana. By the time of maximum volcanism in the Absarokas, and at the time the Great Plains became a site of sedimentation, the Sweetwater Range, although still possessing sharp relief, had sunk to such an extent that it was being covered by shales, tuffs, and sands. This was in Oligocene time. A few remnants of Miocene beds suggest that deposition continued beyond mid-Tertiary time, and certainly the entire range was buried. Then erosion set in, and many of the granite peaks and flanking sedimentary ridges of the old range were reexposed. The stream pattern, as established on the Oligocene and Miocene beds, became superposed on the Precambrian, Paleozoic, and Mesozoic rocks, and the several examples of gorges through the islands are thus explained. The history of burial is detailed under the headings, Hanna Basin and Wind River Basin.

The Sweetwater Range first rose in Fort Union (Paleocene) time and, immediately afterward, was thoroughly eroded during early Eocene, was partly buried by the Wind River beds, and then sank appreciably in late Eocene time.

The islands are in three rows today, the northern reflecting several northwestward-trending anticlines and synclines in the Paleozoic and Mesozoic rocks of the north flank of the range (see *Tectonic Map of the United States*), the central all in the Precambrian core, and the southern revealing southward overthrusting of the Precambrian rocks over the sedimentaries.

WIND RIVER BASIN

The Wind River basin rests between the Wind River and Sweetwater ranges on the southwest and south, and the Absaroka, Owl Creek, and Big Horn ranges on the north. The basin is sometimes construed physiographically to cover the former site of the Sweetwater Range because of the low relief there.

Details of the basin are best known from the work of Love (1939) at the west end, Tourtelot and Nace (1946) at the northeast side, and Van Houten (1957) on the south side. Love's work has already been summarized in connection with the Absaroka Range. The Tertiary formations of the basin are shown in the chart of Fig. 24.5. They range in age from Paleocene to Oligocene, and in parts of the basin they may be over 10,000 feet thick. The basin is asymmetrical with the axis near the north margin. The two chief structural variations from gentle basinward dips in the

eastern part are the Cedar Ridge fault and the McComb anticline. The fault trends northwestward, and the northeast side is down about 1000 feet. It cuts the youngest rocks in the area, and is therefore post-Oligocene.

The McComb anticline is a complex structure and is associated with southward thrusting of Copper Mountain. Thrusting is also indicated in connection with the Cedar Ridge fault. According to Tourtelot and Nace:

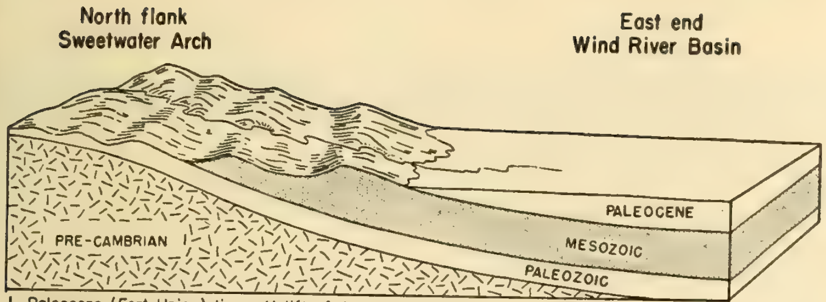
> At the west end of Cedar Ridge, and on the south side of the Cedar Ridge fault, over 1,000 feet of Upper Cretaceous beds, nearly vertical or slightly overturned away from the Big Horn Mountains, are overlain by a thick sequence of boulder beds in the Lysite member of the Wind River formation. The overturning of the Upper Cretaceous beds may be explained by the passage of a thrust sheet of older rocks from the north over them, or by the presence of a thrust sheet just to the north that did not override the Upper Cretaceous rocks but strongly deformed the beds beyond the point of its farthest advance. In addition, as Love points out, there is not enough room for a normal section between the southward-dipping Paleozoic formations and the overturned Upper Cretaceous beds standing about a mile to the south. It is believed that the boulder beds in Cedar Ridge were derived from a thrust sheet which moved southward from the Big Horn Mountains. Knight has postulated a similar origin for boulder beds of this type in the Crooks Mountain area, where the sole of the thrust mass, from which the boulder beds were derived, is exposed. If the boulder beds in the Wind River formation on Cedar Ridge were deposited as erosion products of a thrust sheet, the thrusting must have occurred in Wasatchian (early Eocene) time. These Wasatchian and also younger rocks were cut by the Cedar Ridge fault during or after Oligocene time.
>
> The sequence of major diastrophic events that affected rocks in the northeastern part of the Wind River Basin is summarized as follows:
>
> 1. Mountain building during or at the end of late Cretaceous time.
> 2. Thrust faulting from the north in Wasatchian time along the southern margin of the Big Horn Mountains and the south side of the Owl Creek Mountains.
> 3. Localized gentle folding after the close of Bridgerian time along the southern margin of the Big Horn Mountains.
> 4. Normal faulting during or after Oligocene time along the south end of the Big Horn Mountains and the south side of the Owl Creek Mountains.

The Cenozoic history of the north flank of the Sweetwater Range and the south flank of the Wind River basin is portrayed in a series of block diagrams by S. H. Knight, reproduced in Fig. 24.13.

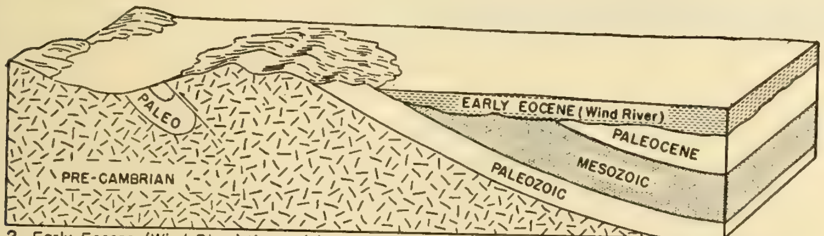
HANNA BASIN

The Hanna basin is bounded on the west by the Rawlins uplift, the north by the Sweetwater uplift, the south by the Medicine Bow Range, but on the east it merges with the northwest end of the Laramie basin. Between the Laramie basin and the Hanna basin is the Carbon basin, through which the two were once continuous but are now separated by Laramide anticlines. The Saddleback Hills anticline separates the Carbon basin from the Hanna, and the Medicine Bow and associated anticlines separate the Carbon from the Laramie. These anticlines are rather sharp and extend northerly from the broad north end of the Medicine Bow Range. The Hanna basin is fairly circular and, although not so large as the other basins of Wyoming, it carries a very thick succession of beds. Paleozoic, Mesozoic, and lower Tertiary formations are over 35,000 feet thick, with Upper Cretaceous, Paleocene, and Eocene accounting for most of the accumulation. The succession is very important because it records better than elsewhere the several episodes of deformation in this part of Wyoming. The formations listed by Dobbin, Bowen, and Hoots (1929) are as follows:

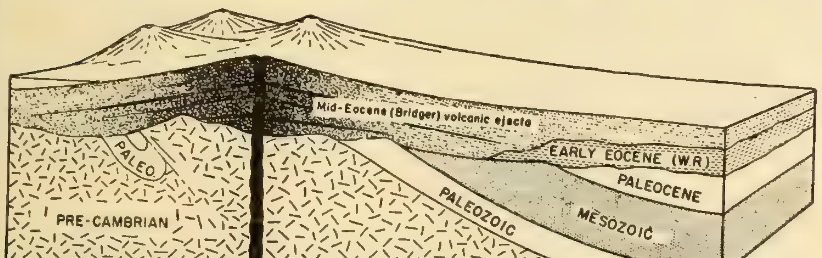
North Park fm. (Miocene?)		0–400 feet
	Unconformity	
Hanna fm. (early Eocene)		7000
	Unconformity	
Ferris fm. (lower part is uppermost Cretaceous)		6500
Medicine Bow fm. (uppermost Cretaceous)		4000–6200
Lewis sh.	Upper Cretaceous	3300
Mesaverde fm.	Upper Cretaceous	2200–2700
Steele sh.	Upper Cretaceous	4000–5000
Niobrara fm.	Upper Cretaceous	700
Carlile sh.	Upper Cretaceous	400
Frontier fm.	Upper Cretaceous	725
Mowry sh.	Upper Cretaceous	120
Thermopolis sh.	Upper Cretaceous	180
Cloverly fm. (Lower Cretaceous)		128
Morrison fm. (Upper? Jurassic)		350
Chugwater (Triassic)		1300
Embar(?) fm. (Permian)		150
Tensleep (Pennsylvanian)		250
Probably pre-Pennsylvanian beds		?



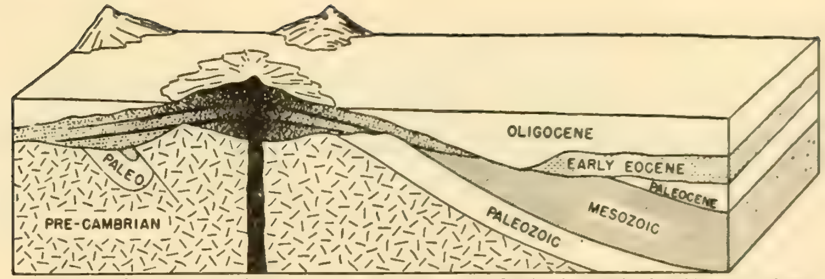
1. Paleocene (Fort Union) time. Uplift of the Sweetwater arch; erosion of the crest of the arch and deposition (Fort Union) in the Wind River Basin to the north.



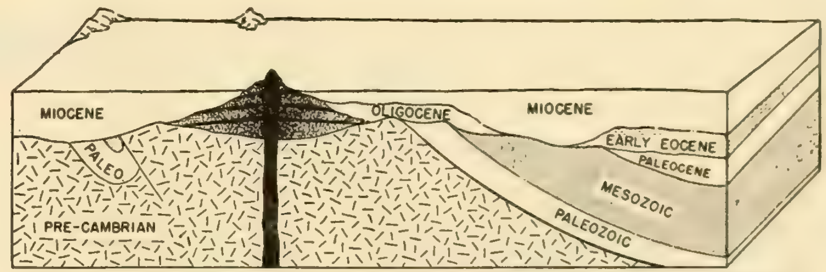
2. Early Eocene (Wind River) time. (1) Development of the thrust faulted Rattlesnake anticline, (2) erosion exposing pre-Cambrian rocks in the core of the anticline, and (3) deposition of Wind River sediments in the topographic lows.



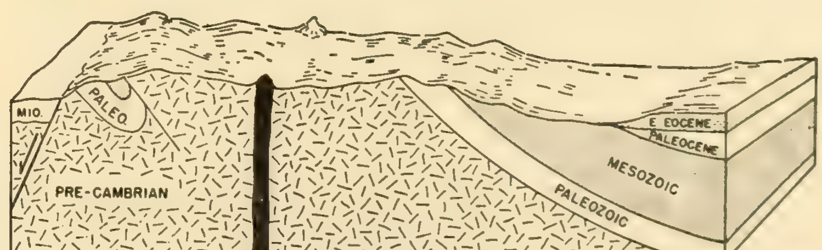
3. Mid-Eocene (Bridger) time. Extensive volcanic activity during which ejecta (mostly fragmental) was expelled through more than a score of vents located along and south of the axis of the Rattlesnake anticline. At the cessation of volcanic activity the ejecta must have been several hundred feet thick in its central portion and probably covered an area of 10,000 or more square miles.



4. Oligocene (late Chadronian) time. Following the cessation of volcanic activity in mid-Eocene time erosion removed much of the volcanic ejecta during late Eocene time. In early Oligocene time the region was buried for the most part or entirely by tuffs. It is believed that these tuffs came from a remote source, possibly from the Yellowstone Plateau-Absaraka Mountains area.

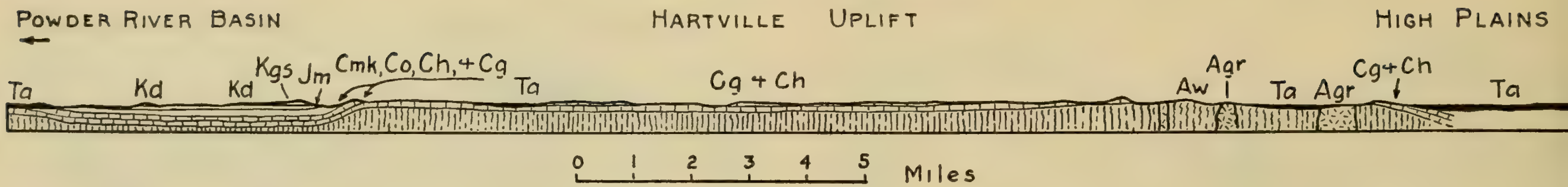


5. Mid-Miocene (Hemingfordian) time. Erosion during late Oligocene (post-Chadronian) time and probably early Miocene (Arikareean) time removed much of the Oligocene deposits and further reduced the remnants of the Rattlesnake ejecta. The region was again buried under deposits of volcanic ashes, sandstones, caliches and conglomerates, which were laid down during mid-Miocene time. Again these deposits came from a remote source.



6. Recent time. Post mid-Miocene movements caused the collapse of the Sweetwater arch although the region as a whole was uplifted. Pliocene and Pleistocene erosion has removed much of the Miocene rocks, reduced the Rattlesnake ejecta to necks and modified the older rocks.

Fig. 24.13. Idealized evolution of north flank, Sweetwater arch. Reproduced from Knight, 1954.



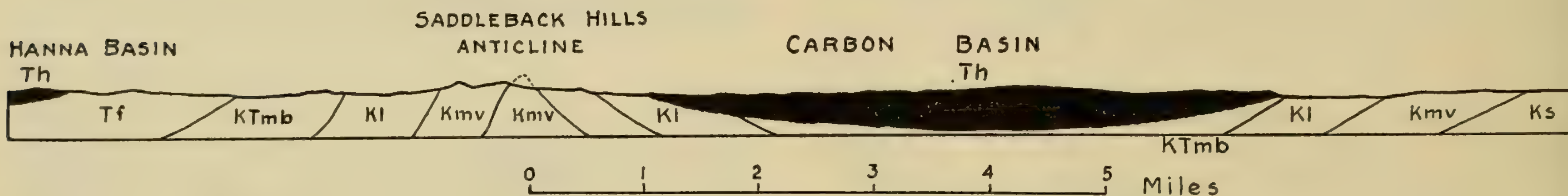


Fig. 24.14. Upper diagram, cross section of the Hartville uplift from the Powder River basin southeastward. Aw, Algonkian Walen group; Agr, granite intrusive into Walen group; Cg, Guernsey fm.; Ch, Hartville fm.; Co, Opeche red ss.; Cmk, Minnekahta ls.; Jm, Spearfish (?), Sundance, and Morrison fms.; Kd, Dakota ss.; Kgs, Graneros sh.; Ta, Miocene Arikaree fm. After W. S. T. Smith, 1903. Lower diagram, cross section from the Hanna basin to the Carbon basin, after Dobbin, Bowen, and Hoots, 1929. Ks, Steel sh.; Kmv, Mesaverde fm.; Kl, Lewis sh.; Ktmb, Medicine Bow fm.; Tf, Ferris fm.; Th, Hanna fm.

The lower cross section of Fig. 24.14 shows the unconformity at the base of the Hanna formation, which, according to Dobbin *et al.* (1929):

> . . . occupies the central portion of the Hanna and Carbon Basins and contains most of the coal mines in this area. It rests uncomfortably on the Ferris formation and transgresses across all underlying formations at least down to the Cloverly and possibly down to the granite. It consists of alternating conglomerate, sandstone, shale, and coal beds, and its base is marked by a thick conglomeratic sandstone and locally by massive conglomerate.

The pebbles of the conglomerate are abundantly of Precambrian derivation, and formations from Tensleep to Mesaverde are also represented. According to the map and cross sections of Dobbin's report, the principal folds of the basin were formed in post-Ferris and pre-Hanna time, and then accentuated in post-Hanna time when the Hanna formation was appreciably folded. But during the first episode of folding, the Medicine Bow Range was vigorously uplifted and considerably eroded to furnish much of the debris for the basin. The Medicine Bow Range had been gently uplifted in Pierre time, and the Precambrian may have been cut into at that early date, but certainly it was widely exposed after the second uplift and during the deposition of the Hanna formation.

The Cenozoic history of the north flank of the Hanna basin and south flank of the Sweetwater Range, together with the thrust structure, is vividly shown in block diagrams by S. H. Knight and reproduced in Fig. 24.15. His comments are are follows:

> The upper diagram represents an early stage in the deformation of the Basin and depicts conditions as they are believed to have existed during early Paleocene time. Erosion has breached the Paleozoic and Mesozoic rock successions and Precambrian rocks are exposed along the crest of the rising Sweetwater Arch. Rock debris derived from the entire succession is accumulating on the shallow Basin floor. Just when, in terms of the local sequence of formations, the central portion of the Sweetwater Arch rose above base level and erosion began to feed debris into the Basin is a question which still remains to be answered. The writer subscribes to the concept that the Sweetwater Arch and the Medicine Bow Mountains may have risen as islands out of the

Cretaceous Sea and that some of the last of the marine deposits, such as the basal Medicine Bow and Lewis, and even possibly the Mesaverde, may be locally derived.

Diagram No. 2, represents conditions as they may have existed during the deposition of the late Lower Eocene (Wind River?). The Paleocene-Eocene (?) succession has been steeply upturned adjacent to the highlands and the large thrust fault has brought the Precambrian in contact with these rocks and they have been truncated by erosion in the vicinity of the highlands. The coarse conglomerates, derived chiefly from the Precambrian, lie with marked angular discordance upon older rocks along the margin of the Basin. These late Lower Eocene (Wind River?) rocks become finer textured and the pronounced angular discordance between them and the underlying rocks disappears as they are traced basinward.

Following the deposition of the Eocene (Wind River?) rocks the region was subjected to moderate folding.

It is apparent that if any Oligocene rocks were laid down in the Basin they were largely or entirely removed before the deposition of mid-Miocene rocks. To the south mid-Miocene rocks rest unconformably upon Cretaceous and older rocks. Following the deposition of the mid-Miocene (Browns Park) the region was subjected to considerable disturbance. A notable feature of this disturbance was wide-spread normal faulting. Available evidence indicates that this disturbance took place in the late Miocene time. It is believed that the region suffered rather extensive uplift during this disturbance. It is probable that the numerous normal faults common to the Basin were formed at this time. Regional evidence indicates that the area was blanketed with sediments during early Pliocene time. The question of the time of the superposition of the North Platte River across the Basin and elsewhere has interested the writer for many years. Until evidence to the contrary is forthcoming, it is concluded that the present course of the North Platte River was established upon the Lower Pliocene surface following regional uplift with some tilting. This uplift began the present cycle of erosion.

LATE TERTIARY DOWNFAULTING OF SWEETWATER RANGE

Along the north and south margins of the exposed Precambrian core of the Laramide Sweetwater Range normal faults of Late Tertiary age have been recognized which have resulted in the downward displacement of the core area some 2500 to 3000 feet (Blackstone, 1951). See Fig. 24.12. Near the volcanic necks of the Rattlesnake anticline the normal faults can be dated as post-middle Miocene.

Since the Late Tertiary downfaulting follows approximately the same pattern as the late Eocene, Oligocene, and early Miocene sagging or

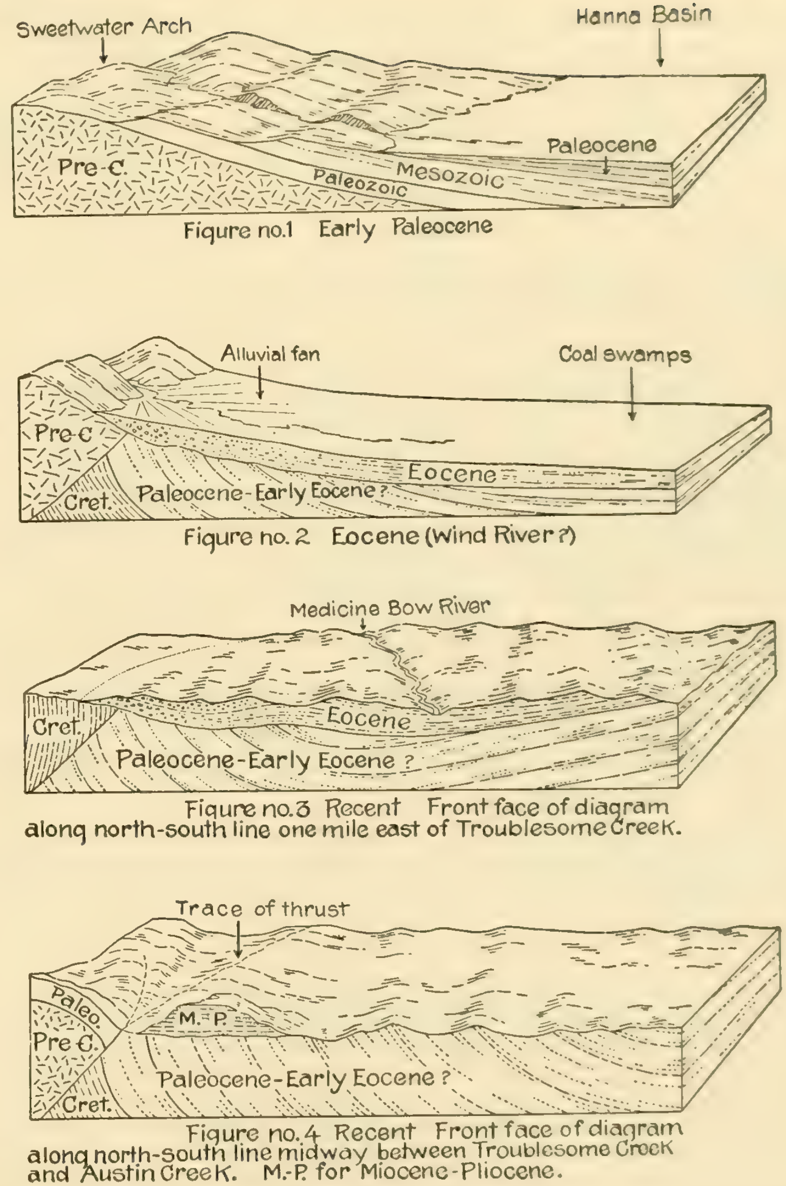


Fig. 24.15. Development of the north flank of the Hanna Basin. Reproduced from Knight, 1951.

downwarping, we may assume that the normal faults are the late result of the subsiding process. In the early stages gentle flexing must have occurred, but later ruptures broke along the sides of the subsiding block.

Tertiary faults have been recognized in several places in central Wyoming, but as yet the extent of the system has not been very well delimited. The pattern of subsidence is outlined under the next heading.

LARAMIDE PATTERN AND CENOZOIC STAGES IN THE SWEETWATER RANGE REGION

It is evident by inspection of the *Geological Map of Wyoming*, from which Fig. 24.12 is taken, that the structures trend in two directions, northwesterly and westerly. From this it may be concluded that two phases of Laramide orogeny occurred. The relations of Casper Mountain to the Emigrant Gap anticline might be taken to indicate that the northwesterly trending structures are the older, and that the east-west trending structures have been superposed on the northwesterly. However, on stratigraphic grounds Love (1954) lists the following succession of events:

1. At the close of Cretaceous the broad Sweetwater Range arch rose.
2. The arch continued to rise during the Paleocene.
3. At the close of the earliest Eocene the thrusting along the south flank of the Sweetwater Range occurred.
4. At the close of Oligocene time the gentle northwesterly trending folds of central Wyoming developed. The settling of the Sweetwater Range was taking place.
5. In post-middle Pliocene and pre-Pleistocene time:

> large-scale block faults developed in many parts of Wyoming; the floor of Jackson Hole dropped several thousand feet; the southern end of the Wind River Mountains collapsed; the central arch of the Granite Mountains (Sweetwater Ridge) dropped several thousand feet; local areas west of the east margins of the Sierra Madre, Medicine Bow, and Laramie mountains were down-dropped; part of the Rawlins uplift collapsed and a broad west-trending anticline formed south of Rawlins; a large area southeast of the Hartville uplift was downfaulted; the southern end of the Big Horn Mountains probably collapsed at this time (Love, 1954).

RAWLINS UPLIFT

The Rawlins uplift (see *Tectonic Map of the United States*) is one of fairly sharp and high structural relief. It stands perhaps 40,000 feet above the crystalline floor of the Hanna basin. It is complicated by folds and thrust faults, and one of the thrusts has brought Precambrian rock in contact with the Mesaverde. This thrust is located just west of the city of Rawlins, and the movement of the overriding sheet is toward the west and southwest. The uplift dips generally northward, toward the Sweetwater uplift, which has been thrust southward against and over it.

WASHAKIE BASIN

Bradley (1945) has reported on the Washakie basin, and his summary is as follows:

> The synclinal structure of the Washakie Basin has given rise to a bold, outward-facing, encircling escarpment, developed on beds in the Green River formation that are more resistant to erosion than other beds in the section. Along the northern margin of the basin this rim rises 600 to 700 feet above the country to the north, and is known as Laney Rim. Southward the escarpment increases in height, and locally on each side of the basin its crest stands about 1,200 feet above the surrounding terrane. Along the southwestern margin of the basin the rim is known as the Kinney Rim. The southward facing escarpment at the southern margin of the basin is broken by stream valleys at many places, and is generally lower. Near the head of Powder Wash, just south of the Wyoming-Colorado boundary line, where the escarpment is low, there is a narrow, southward extension of the Basin, which is clearly shown by the outcrop pattern of the Green River formation. This panhandle is expressed topographically by a pair of outward-facing escarpments that rise to an altitude of more than 8,000 feet at Lookout Mountain.
>
> The rocks in the Washakie Basin are divided into four main units, from bottom to top, the Wasatch, Green River, and Bridger formations of Eocene age, and the Browns Park formation of probable Miocene age. In the broadest and simplest terms, the Green River formation is a huge lens of relatively fine grained fluviatile sandy mudstone that formerly filled a huge intermontane basin far larger than the Washakie Basin. The mudstone is divided into two formations: (1) the Wasatch formation, below the lens of Green River formation, and (2) the Bridger formation above. The sedimentary history of the intermontane basin was complicated, however, by changes in the level of the lake, which resulted in an intertonguing relationship between the Wasatch and

the overlying Green River as shown in the generalized columnar section [Fig. 24.5].

The Washakie Basin is a shallow syncline lying on the east side of the Rock Springs uplift. Along the north and east sides of the basin, the dip of the beds ranges from 3° to 5° toward the center, whereas along the west and southwest sides it ranges from 8° to 12°. In the large central area of the basin the rocks lie nearly flat. This essentially uniform synclinal structure is broken only along the Wyoming-Colorado line, where a west-trending fault zone forms the southern boundary of the basin. This zone, which appears to be an eastward extension of the structural lines in the Hiawatha gas field, is described in an earlier report. The faults of this zone fall into two broad groups, according to their general direction of strike. The faults of the dominant group strike generally westward, whereas, those of the other group strike more nearly northward or northwestward. The faults of the first group are, in general, the older as they are cut by faults of the second group. Moreover, the faults of the second group cut beds of the Browns Park formation, whereas the other faults in most places do not. However, some of the larger east-west faults were apparently active during the second stage of faulting, because locally, as along Cherokee Ridge, they also cut the Browns Park formation.

The rocks in the fault zone are folded into several synclines, the axes of which are parallel to the dominant, westward-striking faults. Most of the folds are rather gentle, having dips that range from 3° to 7°, though locally, as in the vicinity of Baggs, the beds dip as much as 16°.

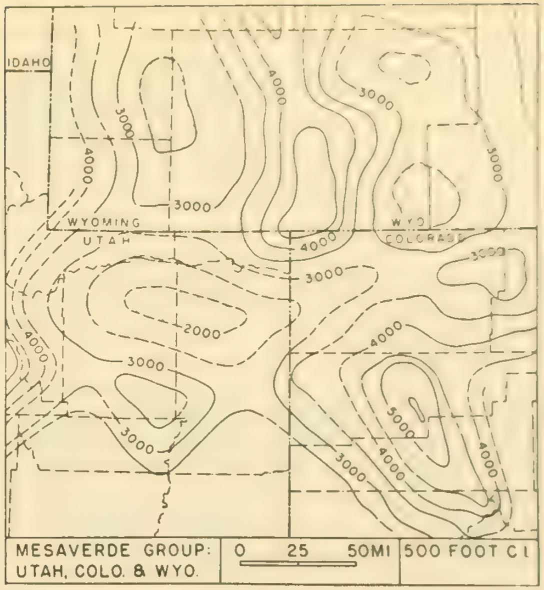


Fig. 24.16. Isopach map of Mesaverde group, Wyoming, Utah, and Colorado. Courtesy John Burger. The site of the present Rock Springs uplift was a basin in Mesaverde time.

GREEN RIVER BASIN

The Green River basin (also referred to as the Bridger basin) is bounded on the south by the great Uinta anticline, on the west by the central Rockies of western Wyoming, on the east by the Rock Springs uplift, and on the north and northeast by the Wind River and Gros Ventre ranges. The extreme northern end of the basin is a wedge between the southwestward thrust Gros Ventre Range and the eastward thrust Hoback Range, and is drained by the Hoback River, a tributary of the Snake River, and hence locally known as the Hoback basin. The Green River drains the rest of the Green River basin. Refer to *Wyoming Geological Association Guidebook*, Tenth Ann. Field Conference, 1955.

The evolution of the Green River basin has been depicted in Chapter 22. See Figs. 22.4 to 22.6 and 24.16. The relation of the Uinta anticline to the basin is shown in Fig. 24.17. From the diagram it may be seen that considerable arching and erosion preceded the deposition of the Green River formation, but from other locations along the north flank of the Uintas the major and sharp rise, involving high-angle thrusting, followed the Green River. Regarding the graded surfaces exhibited on the north flank of the Uinta Mountains and extending far out into the Green River Basin Bradley says:

> Long, narrow remnants of four old erosion surfaces slope gently northward from the north flank of the Uinta Range and truncate the upturned edges of hard and soft beds. The Gilbert Peak erosion surface, which is the highest and oldest of these surfaces, once extended from the crest of the range at an altitude of about 13,000 feet to the center of the Green River Basin. Because undisturbed remnants of this surface have gradients ranging from about 400 feet to the mile near the crest of the range to 55 feet to the mile 35 miles out in the basin, because island mounts of limestone rise rather abruptly from it, and because it apparently never had a soil mantle but is covered in most places by conglomerate, this surface is interpreted as a pediment formed in a semiarid or arid climate. At the time the Gilbert Peak surface was cut the Green River Basin was filled to a greater depth than now with Eocene sedimentary rocks. The Gilbert Peak erosion surface truncated these rocks at very low angles and extended northward across them as a continuous plain. On this plain the master

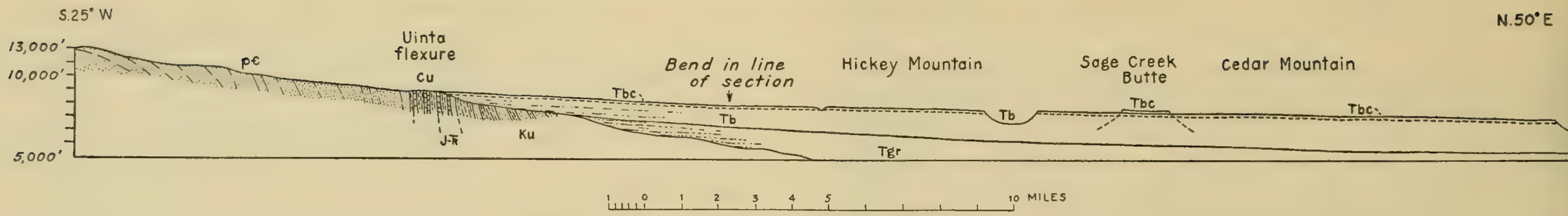


Fig. 24.17. Cross section of the north flank of the Uinta Mountains after Bradley (1936), showing remnants of the Gilbert Peak surface projected laterally to the plane of the section. Note the even truncation of both hard and soft strata. pЄ Uinta Mountain group; Cu, Carboniferous undifferentiated; J–Ꞧ Jurassic and Triassic undifferentiated; Ku, Cretaceous undifferentiated; Tgs, Green River fm.; Tb, Bridger fm.; Tbc, Bishop congl.

stream of the basin apparently flowed eastward to join the ancestral Platte or some similar river that drained into the Gulf of Mexico.

The Bishop conglomerate, which covers much of the Gilbert Peak surface, is coarse-grained and very poorly sorted and fills the deepest concavity in the profile of the pediment, where it is about 200 feet thick. The same streams that cut the Gilbert Peak pediment deposited the Bishop conglomerate, because their transporting capacity changed in response to a climatic shift toward still greater aridity. This climatic change, though critical, probably was not great.

No fossils have been found in the Bishop conglomerate, but the Gilbert Peak surface truncates the latest Eocene rocks and yet is distinctly older than the Browns Park formation (late Miocene or early Pliocene). Hence the Gilbert Peak surface and the Bishop conglomerate are either Miocene or Oligocene. I believe that the Gilbert Peak surface is probably correlative with Blackwelder's Wind River peneplain, near the top of the Wind River Range.

About 400 to 500 feet below the remnants of the Gilbert Peak surface these same streams later cut the less extensive Bear Mountain erosion surface. The characteristics of the Bear Mountain surface are so nearly identical with those of the Gilbert Peak surface that it is regarded as a pediment formed under arid conditions probably closely similar to those which prevailed while the Gilbert Peak surface was being cut. Correlated with the Bear Mountain surface are two large, rather smooth-floored valleys, the Browns Park Valley and Summit Valley. These valleys are in the eastern part of the Uinta Range and are each roughly parallel to the range axis. The floor of the Browns Park Valley descends eastward and passes beneath the Browns Park formation, which is of upper Miocene or lower Pliocene age. As there is no indication that the deposition of the Browns Park formation did not follow immediately the completion of the Bear Mountain surface, that surface is probably also of essentially this geologic age.

After the deposition of the Browns Park formation the east end of the Uinta Mountain arch collapsed by block faulting, . . . and . . . apparently lowered the stream flowing along the ancient Browns Park Valley (on the depositional surface of the Browns Park formation) enough for one of its tributaries, which has already cut through the divide on the north side of the valley, to be rejuvenated and thus to extend its course headward so far northward in the soft Tertiary rocks that it finally captured the ancient master stream of the Green River Basin. When this river, the new Green River, first entered the Browns Park Valley it flowed on the uppermost beds of the Browns Park formation, following the ancient Browns Park stream eastward beyond the east end of the range. But soon thereafter it was captured by Lodore Branch, a tributary to the ancestral Cascade Creek, which drained Summit Valley, and so came to flow along the present site of Lodore Canyon.

UINTA MOUNTAINS

The Uinta Mountains are eroded from a flat-topped anticlinal uplift, the major details in cross section of which are shown in Fig. 24.18. The thrust faulting along the north flank is post-Green River formation and its character suggests horizontal spilling or mass flowage of the margin of the uplift toward the Green River basin as a late or secondary effect of the primary vertical uplift. The vertical uplift of anticlinal crest over basin trough exceeds 32,000 feet.

ROCK SPRINGS UPLIFT

Separating the Green River basin on the west and the Washakie basin on the east is the Rock Springs uplift, a 40-mile-long, doubly plunging, north-south-trending anticline. The resistant sandstones hold up hog-

backs on the west side with dips of 30 degrees, and cuestas on the east side where dips do not exceed 10 degrees. The principal ridge-making sandstone is the Mesaverde, which stands 1000 feet high in places and surrounds the elliptical Baxter basin in the center.

The evolution of the Rock Springs uplift is shown in Figs. 22.3 to 22.6 and 24.16.

The Leucite Hills at the north end of the Rock Springs uplift are remnants of cinder cones and lava sheets. The lavas cap hills of sedimentary rock that now stand 800 to 1200 feet above the plains and preserve remnants of an old, subdued erosion surface (Rich, 1910). This may be equivalent to the Gilbert Peak surface of the Uintas.

The Rock Springs uplift was formed mainly after the Upper Eocene sediments had accumulated. This seems evident because the sediments do not coarsen appreciably toward the uplift. It may have started to rise in late Washakie time incident to the bold arching of the Uintas in modern form, because the upper part of the Washakie formation in the Washakie basin is not present in the Green River basin; but the main elevation of the Rock Springs uplift was post-Washakie.

In the Gilbert Peak erosion cycle of the Uinta Mountains the Rock Springs uplift was beveled, so the folding predated the erosion cycle which terminated in Miocene time.

LARAMIE RANGE AND BASIN AND MEDICINE BOW RANGE

The Laramie basin is, in general, a northward-plunging syncline between the Laramie Range on the east and the Medicine Bow Range on the west. The ranges on either side are formed of Precambrian crystalline rocks, and about 8000 feet of Carboniferous and Mesozoic beds overlie the crystallines in the basin. The sediments are preponderantly shaly. Along the west side of the basin are four anticlines in *en echelon* arrangement, with exposed Precambrian cores; and they are known from south to north as Bull, Ring, Jelm, and Sheep mountains. High-angle thrust faults occur on one or both sides of these anticlines.

Thrust faulting was the chief activity in Laramide times, with the sides of the basin generally bounded by thrusts dipping under the moun-

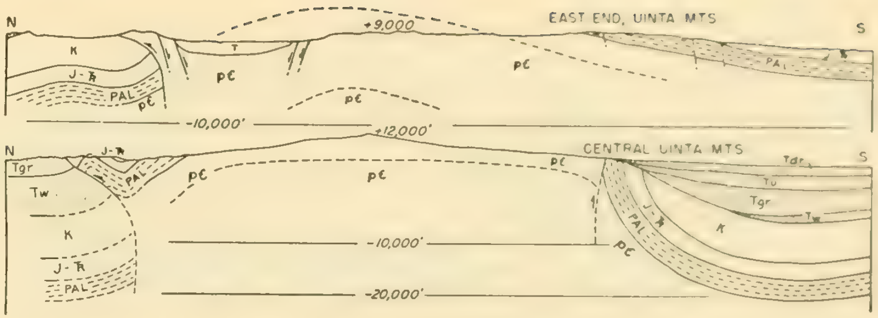


Fig. 24.18. Cross sections of the eastern and central parts of the Uinta uplift. After Ritzma, 1959.

tains. The Medicine Bow Mountains are thrust moderately eastward over the western margin of the Laramie basin (Beckwith, 1938, 1942), and the Front Range crystallines at the southern end of the basin are thrust westward over the sedimentaries. See cross sections of Fig. 24.19. Tear faults and faults that turn into stratification faults at depth without passing into the Precambrian have been described. Younger beds have been thrust over older in places, and at the very south end of the valley, where the thrust sheet from the west is opposed to the thrust sheet from the east, the basin is not as wide as the amount of movement on the thrust surface. From this, Beckwith (1942) concludes that the fronts of the thrust sheets were eroded back sufficiently fast as they advanced so that they did not meet head-on.

The date or phases of deformation cannot be directly determined in the Laramie basin except that they occurred in the interval post-Mesaverde and pre-Oligocene. From reference to the orogenies in nearby Hanna basin and North Park, Beckwith infers that arching of the Medicine Bow and Park ranges started in Late Cretaceous time, while the Cretaceous seas still persisted a short distance from the present mountains. Folding and thrusting occurred during early Eocene and then again some compression shortly afterward cast the lower Eocene beds into folds. The evidence is set forth as follows by Beckwith:

> The folded sediments in the upper Laramie River Valley are about 8000 feet thick. Farther north the Mesaverde is succeeded conformably by 3000 feet of

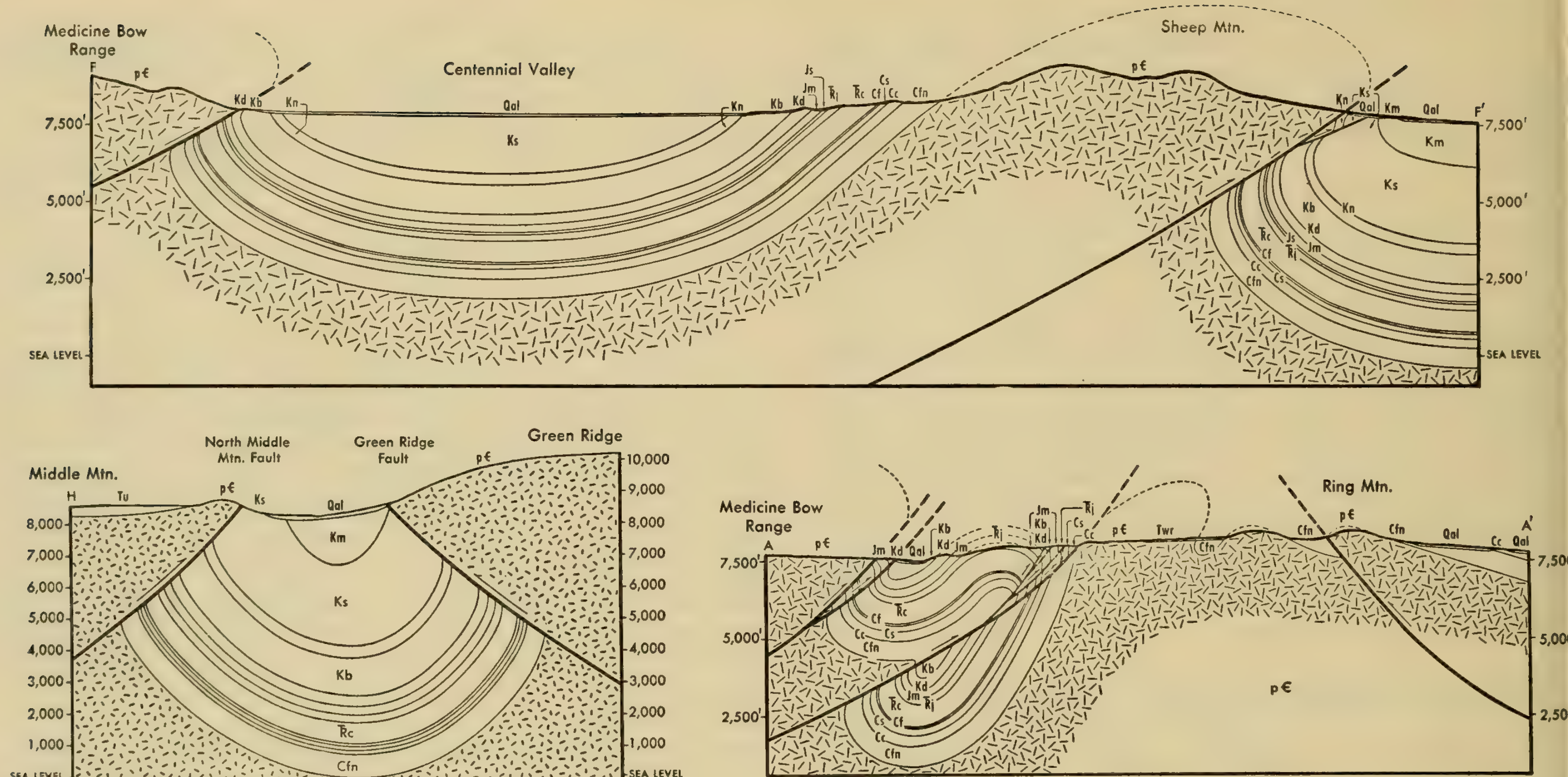


Fig. 24.19. Cross sections along the east front of the Medicine Bow Range and the west side of the Laramie basin, after Beckwith, 1938 and 1942. Cfn, Fountain fm., Cc, Casper fm.; Cs, Satanka sh.; Cf, Forelle ls.; Ꞓc, Chugwater fm.; Ꞓj, Jelm fm.; Js, Sundance fm.; Jm, Morrison fm.; Kd, Dakota group; Kb, Benton group; Kn, Miobara fm.; Ks, Steel sh.; Km, Mesaverde fm.; Kl, Lewis sh.; Twr, White River group.

marine Lewis shale and several thousand feet of grits, standstones, carbonaceous shales, and coals constituting the Medicine Bow formation. The Medicine Bow is overlain unconformably by the Hanna formation. At the Citizen's Coal Mine, 5 miles north of Sheep Mountain, the lower beds of the Medicine Bow contain conglomerates with pebbles of Dakota sandstone and Mowry shale several inches across. Thousands of feet of marine beds must therefore have been stripped from the adjacent rising arch by early Medicine Bow time. A similar conclusion is reached for the region to the south.

Lovering (1935) states:

Before the end of Pierre time, the central part of the Front Range highland was pushed above the level of the sea, and recently deposited shales were exposed to erosion. They were reworked into the upper part of the marine Cretaceous, and the Dakota sandstone was also exposed and reworked in many places and was probably the source of much of the sandy material found in the Fox Hills sandstone.

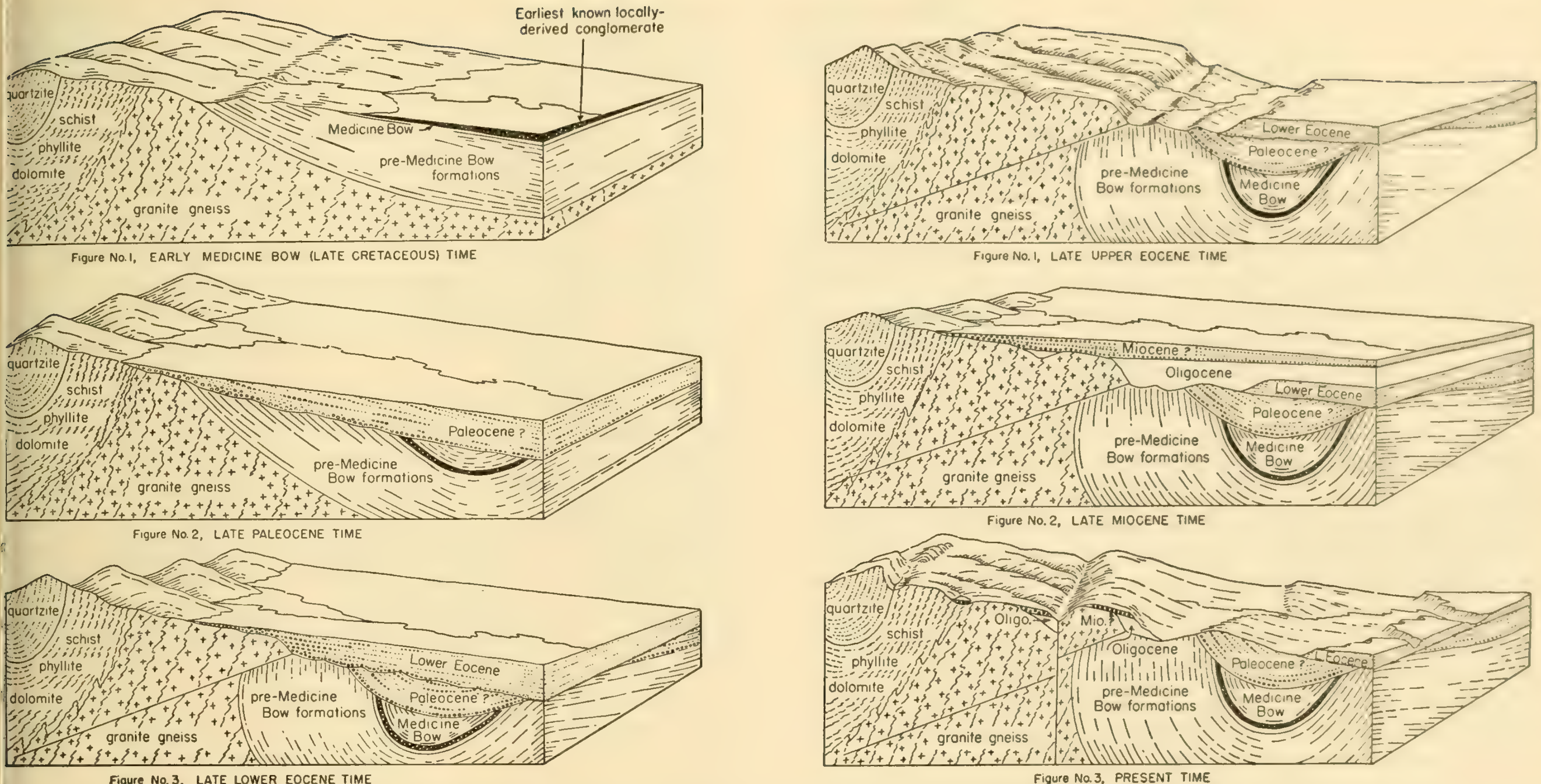


Fig. 24.20. Idealized evolution of Medicine Bow Mountains, Wyo. Reproduced from Knight, 1953.

At the north end of the Laramie basin is a belt of folds arranged *en echelon.* The belt is on the easterly projection of the Sweetwater uplift, and the direction of the fold axes in the *en echelon* belt is northeasterly. Jurassic and Triassic rock is exposed in the cores of the anticlines, and on the north they are blanketed with the Oligocene White River beds.

The evolution of the Medicine Bow Mountains in Cenozoic time is shown in Fig. 24.20.

HARTVILLE UPLIFT

The Hartville uplift is a northeast arm of the Laramie Range, and connects it effectively with the Black Hills uplift. The Laramie Range is a broad, flat-topped anticline, or uplifted plateau with bounding monoclinal flexures, and so also is the Hartville uplift, if viewed from the south end of the Powder River basin to the Great Plains. See cross section of Fig. 24.14. The Hartville uplift is broadest at its junction with the Laramie

Range but narrows toward the Black Hills. The east-bounding monocline has a structural relief of about 9000 feet within a few miles, and the one on the northwest drops the strata as abruptly and about an equal amount.

The top of the Hartville uplift is about 2000 feet below that of the Laramie uplift. Its sedimentary veneer has been stripped off only in a narrow zone along the axis; from this it is deduced that the uplift was never as high as the Laramie or Sweetwater uplifts, and that its relatively lower position today is not due principally to late Laramide subsidence like that of the Sweetwater, but to the fact that it was never elevated high enough in the face of much higher elevations nearby to have suffered as much erosion. It was largely buried by the White River beds in lower Oligocene time, and in the later cycles of erosion it has been partly exhumed.

REGIONAL UPLIFT IN LATE CENOZOIC

The Eocene deposits of Wyoming and adjacent areas accumulated in swamps, on flood plains, and in fresh-water lakes. The flora and fauna indicate a warm, rain-forest climate, and the elevation above sea level at which the sediments were laid down is generally considered to have been not in excess of 1000 feet. Today they occur at about 7000 feet, especially in the Green River, Wind River, and Big Horn basins. The Great Plains adjacent to the Rockies contain Early Tertiary sediments deposited at low levels, but which now stand in places as high as 5000 feet. It is patent that uplift on a very broad scale has occurred. We must be aware of the history of subsidence and sedimentation in central Wyoming, the eastern end of the Uintas, and elsewhere in mid and late Eocene, Oligocene, and early Miocene time, and post-Miocene normal faulting resulting in further subsidence. However, in the broad picture from the Northwest Territories of Canada, southward through Alberta, and the outer Rockies of Montana, Wyoming, Colorado, and New Mexico, the dominant late Cenozoic activity was uplift, and in an amount from 2000 to 6000 feet. The southern part of the Colorado Plateau was uplifted perhaps 8000 feet.

Most of the literature concerning the erosional and depositional activity during the building of the Laramide Rockies and the later regional uplift depicts a history as follows. Immediately after the Laramide ranges were uplifted, extensive erosion surfaces were developed—in places several, one below the other—indicating times of crustal stability separated by uplift and dissection. The erosion surfaces in places can be traced out and are said to level the basin fill deposits. Then, with the regional uplift, just mentioned, the erosion surfaces were greatly dissected and produced our present topography.

Mackin, Van Houten, and others more recently view the history as follows. Erosion affected the uplifts immediately as they appeared from the Cretaceous seas and removed sediments to the intervening basins. The end result of the erosional and transportational processes was a vast graded surface, in part erosional and in part depositional. With the building of great volcanic piles in the Yellowstone and Absaroka region in mid- and late Eocene and Oligocene time the streams draining eastward, perhaps fanlike from the volcanic field, were overloaded with fine debris, and according to Love (1956b), the intermontane basins of Wyoming were so filled that the large bordering ranges were almost submerged. Volcanic activity broke out in other areas, and as pointed out some areas subsided, so that sedimentation was of irregular thickness in places and continued where subsidence continued. But in late Miocene or early Pliocene time most of the graded surface was uplifted, an arid climate resulted, and a regimen of erosion started. This has continued in most places until today. With the coming of the arid and semiarid climate the grassy plains came into existence, and many animals evolved and adapted to a life on the open prairie.

25.

COLORADO AND NEW MEXICO ROCKIES

EXTENT OF LARAMIDE DEFORMATION

Most geologists regard the high relief features in central Colorado, composed principally of the Park, Front, Sawatch, Wet, and Sangre de Cristo ranges, as the Laramide structures of the state. The belt is about 80 miles wide and extends in a north-south direction. See Fig. 25.1. It continues southward to southern New Mexico where it joins the Laramide belts of Mexico and southern Arizona.

An arm extends southwestward from the Sawatch Range to and including the San Juan Mountains.

COLORADO ROCKIES

Geography

Figure 25.2 shows the principal features of geologic interest in Colorado, and on it the above-mentioned belt of Laramide deformation may be identified. The Front Range is the largest and highest of any in the Rockies of the western United States. The most rugged mountains are in the central part with a number of peaks exceeding 14,000 feet in elevation. The western flank slopes steeply away from the crest of the range, but the eastern slope is characterized by broad, dissected, benchlike erosion surfaces that descend in steps to the Great Plains.

A series of valleys or basins occupy a central position to the flanking ranges on east and west, namely, North Park, Middle Park, South Park, and Huerfano Park. The San Luis Valley lies west of the Sangre de Cristo Range and continues the basins in offset fashion into New Mexico.

The Colorado Plateau extends across western Colorado to the Park and Sawatch ranges and is generally considered to include the Piceance basin, the White River uplift, and the Uncompahgre uplift. An extensive volcanic field obscures much of the Laramide geology between the Needle Mountains of the San Juans and the Sawatch and Sangre de Cristo ranges.

The Denver, Trinidad, and Raton basins on the east of the Colorado Rockies are Laramide downwarps.

Relation of Laramide Rockies to Ancestral Rockies

The Colorado Range of the Ancestral Rockies has been described on previous pages. It was gradually overlapped from the east and west and nearly buried by late Cretaceous time. During the Laramide unrest the modern Front Range rose approximately along the eastern half of the ancestral range, and the basins of the Middle and North parks and the Park Range appeared approximately along the western half. See Figs. 25.3 and 25.4

The ancestral Central Colorado basin with its thick fill of Pennsylvanian and Permian red beds and evaporites, and also an appreciable thickness

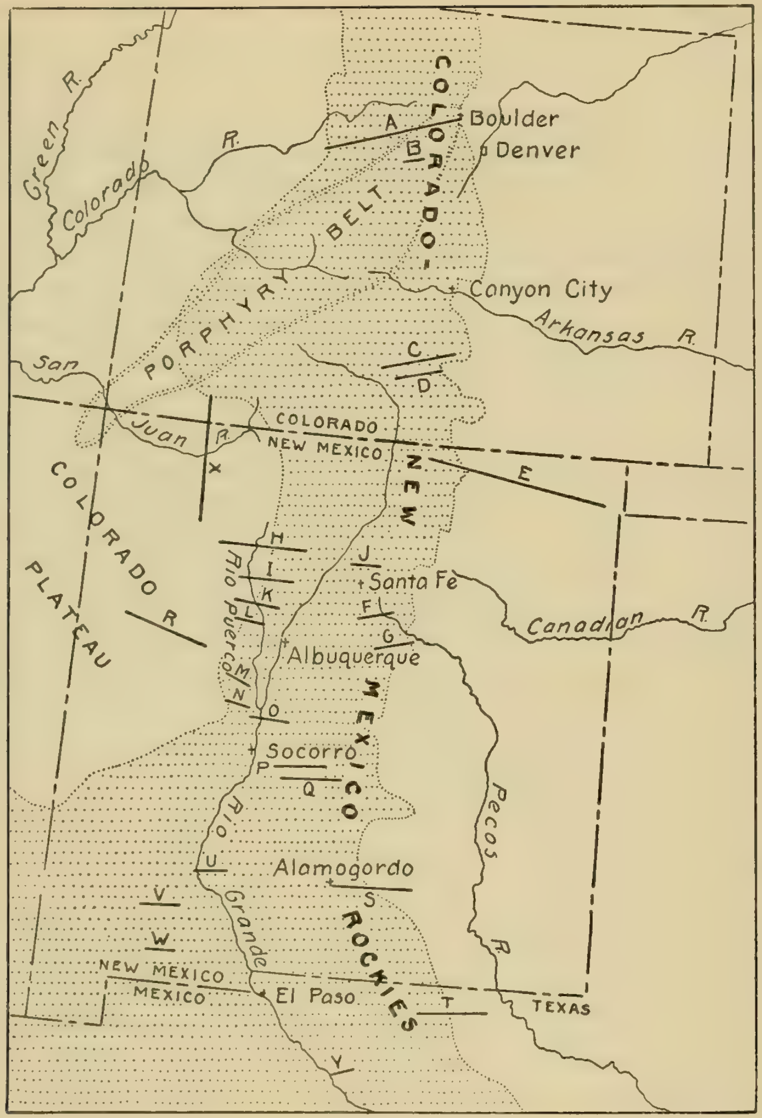


Fig. 25.1. Index map for cross sections of Colorado and New Mexico. Refer to Raisz' *Landforms Map of the United States* for location of ranges, mountains, and valleys. Stippled area is zone of marked Laramide disturbance.

of later Cretaceous shales and sandstones, was considerably deformed in Laramide times. The evaporites contributed to great distortion of the beds in the Eagle area (Fig. 25.4), and the White River uplift or arch as a continuation of the Uinta uplift became the largest structure in the ancestral basin.

Central Parks

Although Middle Park is a well-defined drainage basin, it is more complex topographically and structurally than the other parks. North and South parks have little, or only moderate, internal topographic relief; structurally they are broad, open synclines, marginally faulted, particularly along the eastern sides. Middle Park is also generally synclinal, but the structural and topographic continuity is disturbed by many projecting mountain spurs characterized by overthrust faulting. Northward trending spurs of the Front Range, comprising the Williams River Mountains and the Vasquez Mountains, extend into the southern part of the park as far as the Colorado River. The northern part is largely occupied by spurs projecting southward from the Never Summer Mountains and from the unnamed ridge followed by the continental divide between Middle and North parks.

Front Range

The Front Range is mostly an expanse of Precambrian rock with upturned or overthrust sedimentary rocks on east and west flanks. The Precambrian rocks have been reviewed in Chapter 4. According to Lovering and Goddard (1950):

> The western side of the Front Range is marked by a series of great overthrust faults that formed at this time from the southern end of the South Park as far north as the Wyoming line. The displacement on the Williams Range thrust fault north of Breckenridge is more than 4½ miles, and the movement on the Never Summer thrust north of Branby is more than 6½ miles. The eastern side of the Front Range was subjected to much less severe deformation but was the locus of many echelon northwesterly folds and persistent steep northwesterly faults. Its structure is dominantly that of a steep monoclinal fold, though locally, as at Colorado Springs and Boulder, some thrusting has taken place.

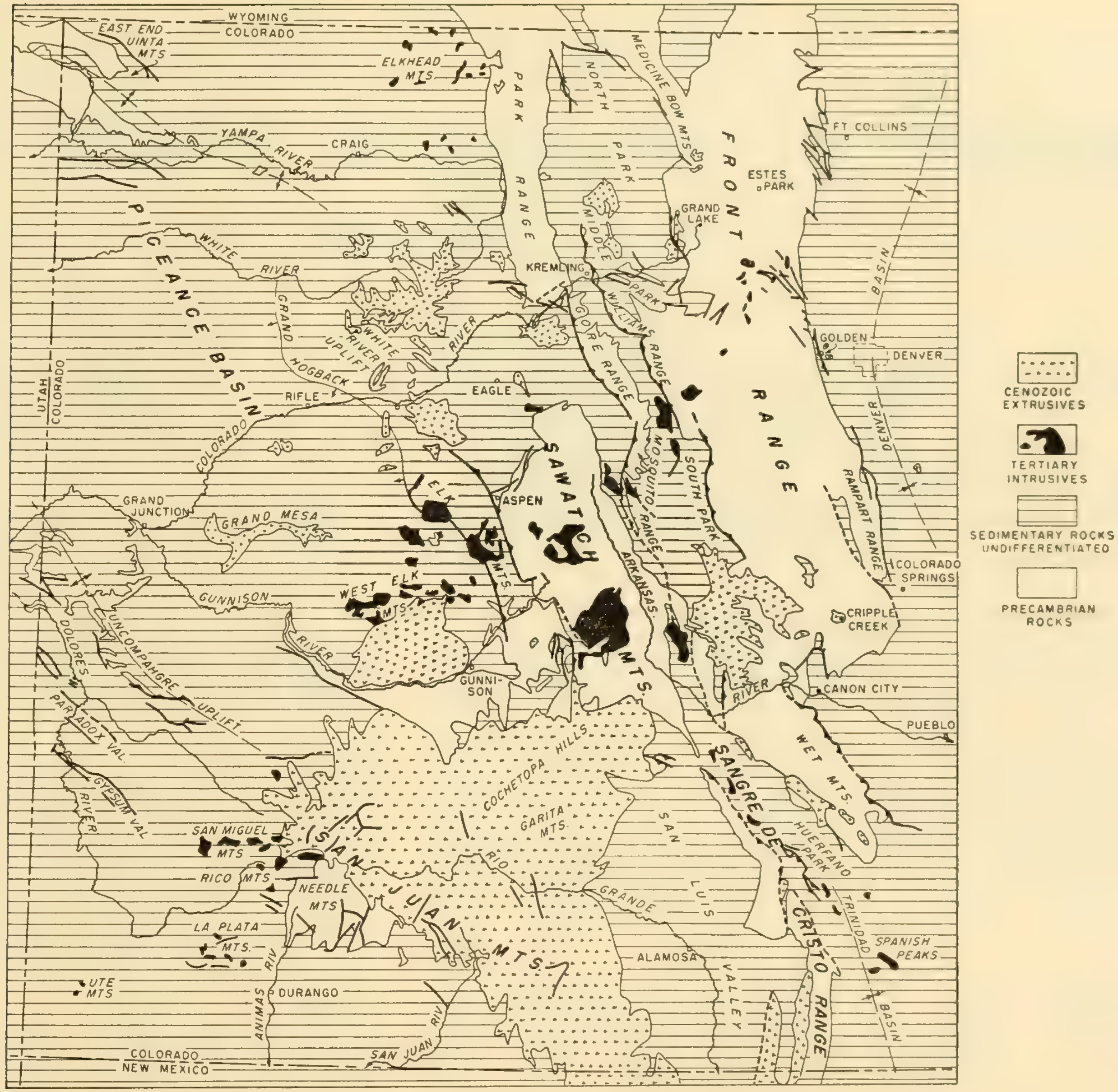


Fig. 25.2. Index map of central and western Colorado.

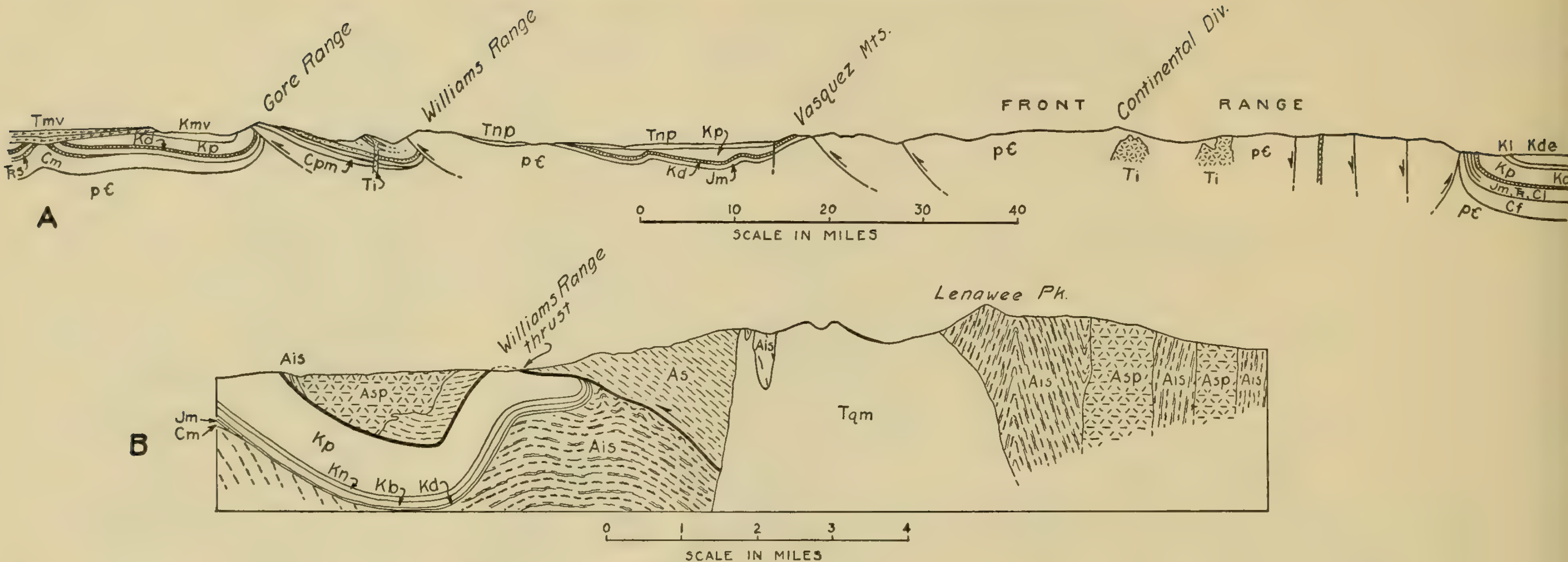


Fig. 25.3. Upper section generalized across the Front Range of Colorado and three of the back ranges, from Boulder to State Bridge. pЄ, Precambrian undifferentiated; Cf, Fountain formation; Cl, Lyons formation; TR Lykins formation; TRs, State Bridge siltstone; Cm, maroon formation; Jm, Morrison formation; Kd, Dakota sandstone; Kp, Benton, Niobrara, and Pierre formations; Kl, Laramie formation; Kde, Denver formation; Kmv, Mesa Verde formation; Ti, Tertiary intrusives; Tnp, North Park fm.; Tmv, Miocene volcanics. Section A of Fig. 25.1.

Lower section is a detail of the Front Range near Montezuma, after Lovering, 1935. Ais, Idaho Springs formation; As, Swandyke hornblende gneiss; Asp, Silver Plume granite; Cm, maroon formation; Jm, Morrison formation; Kd, Dakota sandstone, Kb, Benton shale; Kn, Niobrara formation; Kp, Pierre shale; Tqm, Tertiary intrusives. Section B of Fig. 25.1.

The period of overthrusting was followed by northeasterly and east-northeasterly faulting on a large scale throughout the mineral belt during and after the intrusion of the porphyritic rocks that dot it. Many of the mineral deposits are localized at the intersection of easterly and northeasterly faults with the earlier persistent northwesterly faults where they cross the mineral belt.

Faults formed after the Laramide revolution are comparatively local and largely confined to Miocene volcanic areas and Tertiary basins close to the mountain front.

A group of northeast-trending faults is confined mostly to the western part of the range and seems to mark the western limit of the mineral belt. They appear to be steep and are marked by gougy shear zones 10 to 600 feet wide. The largest of the group is the Moffat Tunnel fault (Lovering and Goddard, 1938a), which is intermittently exposed on the surface for a distance of more than 25 miles and forms a wide zone of "heavy ground" 1000 feet wide in the Moffat Tunnel, 2000 feet below the outcrop. It passes through Berthoud Pass, where the zone of fractured rock is 200 feet wide, and through Loveland Pass, where it is less than 50 feet wide. The badly broken rock is apparently responsible for these depressions.

The Rocky Mountains through central Colorado have generally been regarded as a belt of horizontal compression, but they can also be interpreted as a group of closely packed uplifts with the Front Range longer and wider than any of the others in the entire province. The ancestral Colorado Range was an uplift almost as large as the entire cluster of ranges in the central Colorado belt, but the later Laramide uplifts de-

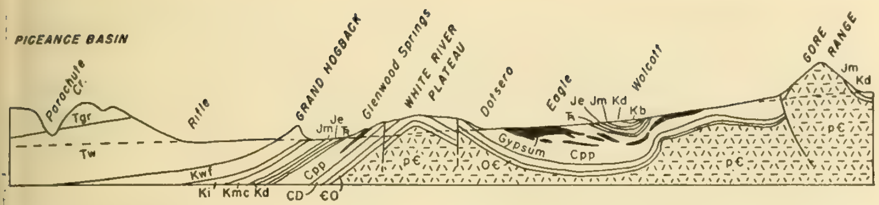


Fig. 25.4. Section along the Colorado River from the Gore Range to Rifle and the Piceance basin. Tgr, Green River fm.; Tw, Wasatch fm.; Kwf, Williams Fork fm.; Ki, Iles fm.; Kmc, Mancos sh.; Kb, Benton gr.; Kd, Dakota ss.; Jm, Morrison fm.; Je, Entrada ss.; ℞, Chinle sh.; Cpp, Permian and Pennsylvanian; CD, Miss. and Dev.; O€, Ordovician and Cambrian. After Bench *et al.*, 1948.

veloped as independent units, not much controlled by the ancestral uplift. The flanks of the central Colorado ranges are replete with thrusts, and apparently reflect the superior uplift of the Front Range.

Harms (1961) has studied the sandstone dikes of the eastern margin of the Front Range south of Denver and presents a convincing case for granite tectonics there. Large Laramide faults place Precambrian rocks in contact with sediments as young as Tertiary in age. The stratigraphic displacement in places is 15,000 feet and the structural relief from 15,000 to 25,000 feet. He concludes that the stress distribution causing the injection of the sandstone dikes was governed by dip-slip movement along steeply westward dipping, convex upward fault surfaces, and that, therefore, the major structures outlining the flank of the range are high-angle reverse faults which steepen with depth.

Uplift of the Front Range began in middle Pierre time while the Denver basin was sill being downwarped, and from that time till well into the Paleocene the central part of the range moved upward at an ever increasing rate. Parts of it rose above the ocean during Fox Hills time, and at the beginning of Denver time large areas were shedding pre-Cambrian debris to the east and west. Intense folding and faulting occurred at the edges of the basins of deposition where the troughs merged with the old positive element about the end of Denver time and outlined the Front Range as it now is (Lovering and Goddard, 1950).

North Park Thrusts

An unusual example of thrusting in the general Front Range region, and as indicated previously one that represents considerable horizontal movement, is at Cameron Pass. Here the Never Summer Range borders on the southern part of the Medicine Bow Mountains adjacent to North Park. A tear fault extending along the Middle Fork of Michigan Creek separates two patterns of thrusting (Gorton, 1953). See map of Fig. 25.5. Although they developed simultaneously, each produced its own structures. The block on the north exhibits two thrusts, as shown in the upper section of Fig. 25.6, whereas the block on the south is interpreted to have one thrust. All thrusts have been folded, and the sequence of events appears to Gorton as follows:

1. Folding, probably in Late Cretaceous.
2. Thrusting, post-Middle Park and pre-North Park.
3. Open folding. The quartz monzonite stock was emplaced during this stage or immediately afterward.

The Benton Gulch thrust klippe and the downfolded Never Summer slice would be interpreted by some geologists as detached gravity slide blocks, and the amount of compressional orogeny minimized. The writer is inclined to view vertical uplift as of paramount importance with mar-

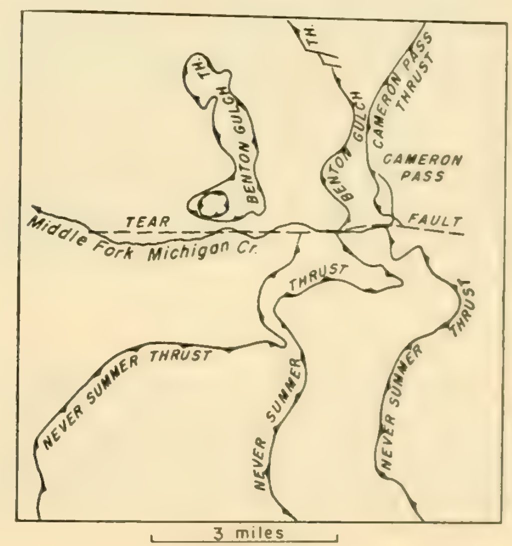


Fig. 25.5. Thrusts of the Cameron Pass area, Never Summer Range, Colorado. After Gorton, 1953.

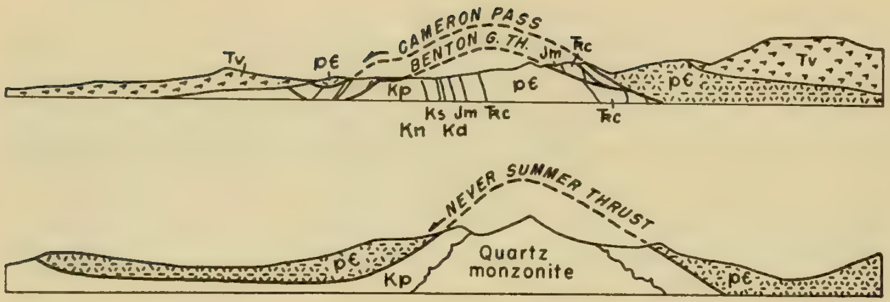


Fig. 25.6. Cameron Pass cross sections. Upper section north of tear fault; lower section south of tear fault. Refer to Fig. 25.5. After Gorton, 1953. TRc, Chugwater fm.; Jm, Morrison and Entrada; Kcl, Dakota ss.; Ks, Benton sh.; Kn, Niobrara ls.; Kp, Pierre sh.

ginal gravity flow movements auxiliary to the vertical in the Colorado and New Mexico Rockies. See discussion under New Mexico Rockies.

A transverse fault, called the Independence Mountain, is prominent at the north end of North Park. It trends N. 65° W., dips at a low angle northward, and has been mapped for 20 miles (Blackstone and de la Montagne, 1960). Precambrian gneisses have been thrust southward over all formations from Triassic Chugwater to Paleocene Coalmont. The overlapping nature of the Coalmont indicates that it was derived from previous uplift, and some subthrust folds suggest previous folding also. The thrust is of Eocene age and is offset by post-late Miocene normal faults. Isolated Precambrian rocks resting on the Coalmont formation south of and at lower elevations than the trace of the thrust are considered by Blackstone and de la Montagne to have reached their position by gravity sliding, but they do not propose gravity sliding for the main thrust sheet.

Transverse Porphyry Belt

Many small instrusive bodies in Colorado will be noted on the map of Fig. 25.2. They are largely concentrated in a narrow diamond-shaped belt that extends from the southwest corner of Colorado, even from the adjacent states of New Mexico, Arizona, and Utah, northeastward through the San Juan Mountains and the Sawatch Range to the Front Range, and across it to Boulder City. They compose the so-called porphyry belt, and most of the state's mineral deposits are localized in it. The belt trends nearly normal to the belt of thrusting. See Fig. 25.1. All the small stocks are Laramide in age and their intrusion generally accompanied the mountain building.

According to Lovering and Goddard (1938b):

> The late Cretaceous and early Eocene (Laramide) igneous rocks or "porphyries" of the mineral belt are readily distinguished from all but a very few of the pre-Cambrian rocks, and many different varieties are so distinctive in appearance as to justify correlation between districts separated by several miles. See Fig. 25.7. These igneous rocks are commonly medium- to fine-grained and nearly all are prophyritic. Some of the early rocks of mafic or intermediate character are holocrystalline, and thin dikes of widely differing composition have felsitic to glassy textures. These intrusives show a wide range in chemical and mineralogic composition, and include dikes as mafic as limburgite, as silicic as alaskite, and as alkalic as aegirite syenite. Most of the intrusive rocks are intermediate or siliceous porphyries whose compositions range from hornblende diorite or biotite-quartz monzonite.
>
> The mineralization followed the intrusion of the Eocene porphyry. The extensive lead-silver deposits of Leadville, the iron-zinc deposits of Gillman, the molybdenite deposits of Climax, the lead-silver deposits of Montezuma, Silver Plume, and Georgetown, the gold deposits of Gilpin County and Central City, and the tungsten deposits of Nederland were formed at this time. Although most of the mineralization in the San Juan Mountains occurred in Miocene time, some deposits of Eocene age were formed near the centers of early Tertiary intrusion in that region.

Sangre De Cristo Range

The Wet Mountains are almost separated from the Front Range by the Canyon City embayment, but in the early history of the Ancestral Rockies and the later history of the Laramide orogeny, the two were closely associated. The eastern border of the Wet Mountains is characterized by Laramide overturning or overthrusting of the Carboniferous and Mesozoic formations toward the east. The western border is one of sharply upturned beds and steep faults.

West of the Front Range in the area of overlap on the ancestral Colorado Range and near the center of the Carboniferous basin, sharp folds and several large thrust faults have resulted in a chain of ranges, principal of which are the Sawatch and Sangre de Cristo. They extend to the

east front of the Rocky Mountains south of the Wet Mountains. See Fig. 25.2. The Sangre de Cristo Range extends southward into New Mexico 110 miles and has a regular arclike form, convex toward the east. Its width is small, ranging from 10 to 20 miles. Together with the San Luis Valley on the west and the Wet Mountain Valley and Huerfano Park on the east, it represents a prong of the Laramide belt. Complex folds and overthrusts dominate the belt. The folds in the northern part of the range consist of a major central anticline bounded on the west by a much compressed, overturned, and faulted syncline, and on the east by a more open syncline. The major anticline is complicated by the presence of a metamorphic core and by several small intrusive bodies along or near its axis. The overthrusts are best preserved in the Huerfano Park region, where several imbricate thrust sheets were thrown into steeply inclined positions by the compressional forces (Burbank and Goddard, 1937). See Fig. 25.8.

The upper section of Fig. 25.9 represents a supposed early stage of compression and overthrusting prior to the upthrust of the Precambrian rocks and the downfaulting of the San Luis Valley; the lower section, the structures afterward. The anticline is viewed as an injective mass due to considerable mobility of the shaly beds of the Lower Pennsylvanian which were overlain by a great thickness of less mobile conglomerates. When compressed, the shales flowed into the core. The belt of plastic deformation is limited to an area just east of the arc-shaped bend in the thrust zone.

The belt of thrusting along the east side of the Sangre de Cristo, opposite Huerfano Park, is illustrated in Fig. 25.8. The principal thrusting occurred after the deposition of the Poison Canyon formation (lowermost Eocene or Paleocene) and before the deposition of the Chuchara formation (middle ? Eocene). See stratigraphic chart of Fig. 25.10. Both the Chuchara and overlying Huerfano formations are affected by the thrusting; but Burbank and Goddard believe the thrusting, although continuing into Eocene time, was of declining intensity.

West of, and inside the arc of thrusting, are two elongate masses of Precambrian rock which were buried or were much lower in elevation than now during the thrusting, but which were later elevated as blocks bounded by high-angle faults. Accompanying the uplifting was considerable plastic deformation of the adjacent shales. The maximum vertical uplift is estimated as 2 to 3 miles (Burbank and Goddard, 1937), and most of it occurred in post-Huerfano (late ? Eocene) time.

After the Precambrian "massifs" were uplifted, they were broken by tensional faults, and in part settled so much as to be covered by effusions

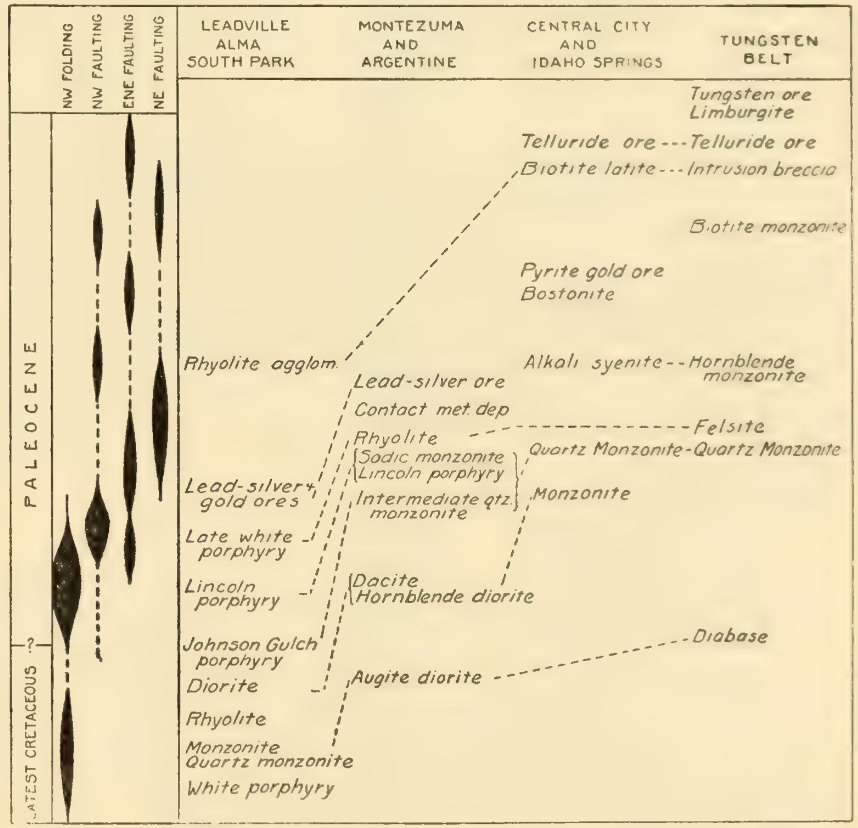


Fig. 25.7. Laramide folding, faulting, and emplacement of igneous rocks and ore deposits in the Front Range mineral belt, Colorado. Abbreviated after Lovering and Goddard (1938b). The dashed lines connect intrusions and ore deposits of similar kind. Equivalent age is portrayed by similar horizontal position. From southwest to northeast the intrusions and ore deposits become generally progressively younger.

Fig. 25.8. Map and structure section of a lobate salient of the main thrust belt, Sangre de Cristo Range, Bluff Creek area, Huerfano Park, Colorado. Reproduced from Burbank and Goddard, 1937.

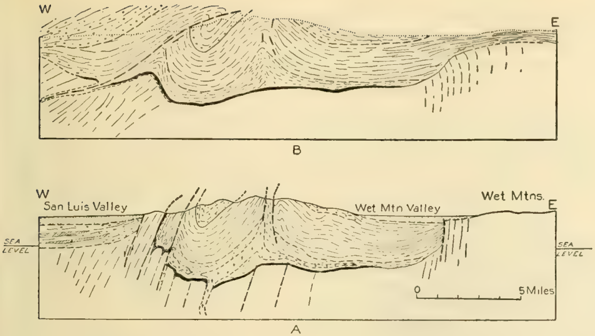


Fig. 25.9. Generalized and diagrammatic sections across the north central part of the Sangre de Cristo Mountains. See section C, Fig. 25.1.

Section A constructed chiefly from a traverse across the range near Crestone. Symbols are generalized: solid black pattern (base) represents lower Paleozoic sandstones and limestones; overlying crinkled line pattern, Lower Pennsylvanian beds involved in zones of shearing and plastic deformation; lighter line and conglomerate pattern, higher Pennsylvanian and Permian formations; the filling of the San Luis Valley depression is Late Tertiary and Recent alluvium, and older lavas and gravels; intrusive bodies are entirely hypothetical but are at positions corresponding to similar bodies exposed in other parts of the range.

Section B is a diagrammatic representation of the same section, based upon the hypothesis of an earlier phase of lower-angle overthrusts, which are presumed to be responsible for the greater part of the shallower tangential deformation in the marginal belt. From Fig. 1, Burbank and Goddard, 1937.

PERIOD	FORMATION			THICKNESS IN FEET	GENERAL CHARACTER
PLIOCENE OR PLEISTOCENE	Terrace gravels				
EOCENE	Huerfano fm.			2300-3500	Marls, clays, soft sandstones and shales, predominantly red, but in part gray, yellow green and purple.
	Cuchara fm.			300-500	White and pink sandstone with thin layers of shale; surface cavernous.
	Poison Canyon fm. Possibly including beds equivalent to Raton formation.			2000-3500	Arkosic sandstone and fine conglomerate, with thin beds of yellow clay; lower beds weather pale yellow.
UPPER CRETACEOUS	Vermejo fm.			0-450	Dark shale, light gray friable sandstone and coal in alternate layers.
	Trinidad ss.			160-225	Massive sandstone, shaly in lower part
	Pierre shale			1800-2000	Yellowish gray to dark gray shales, with zone of impure limestone concretions.
	Apishapa sh.	OF NIOBRARA AGE		450-500	Bluish gray shales at base, grading upward through papery shales to sandy shales
	Timpas ls.			180-200	Grayish white ls and calcareous sh.
	Carlile sh.	OF BENTON AGE		170-180	Dark gray shale, capped by yellowish ss.
	Greenhorn ls.			30-40	Thin-bedded dove colored limestone.
	Graneros sh.			200-310	Gray to black sh. with concretions near base
	Dakota ss.			350-400	Dense, fine grained sandstone
LOWER CRET.	Purgatoire fm.				Coarse, yellowish gray ss., overlain by thin bed of gray shale.
UP. JURASSIC	Morrison fm.			100-240	White sandstone, pink and green shales and hard fine grained limestone.

Fig. 25.10. Mesozoic and Tertiary formations of Huerfano Park and vicinity, Colorado. From Plate 3, Burbank and Goddard, 1937.

of lava. This is a post-Laramide tectonic event, and it is believed to have started not sooner than late Oligocene.

The structure of the Sangre de Cristo Rrange southward in New Mexico is less complicated, and the mountain front resembles that of the Colorado Front Range (Smith and Ray, 1941).

NEW MEXICO ROCKIES

Geography

The Rio Grande flows southward through central New Mexico from Colorado to El Paso, Texas, a distance of 450 miles. It occupies a series

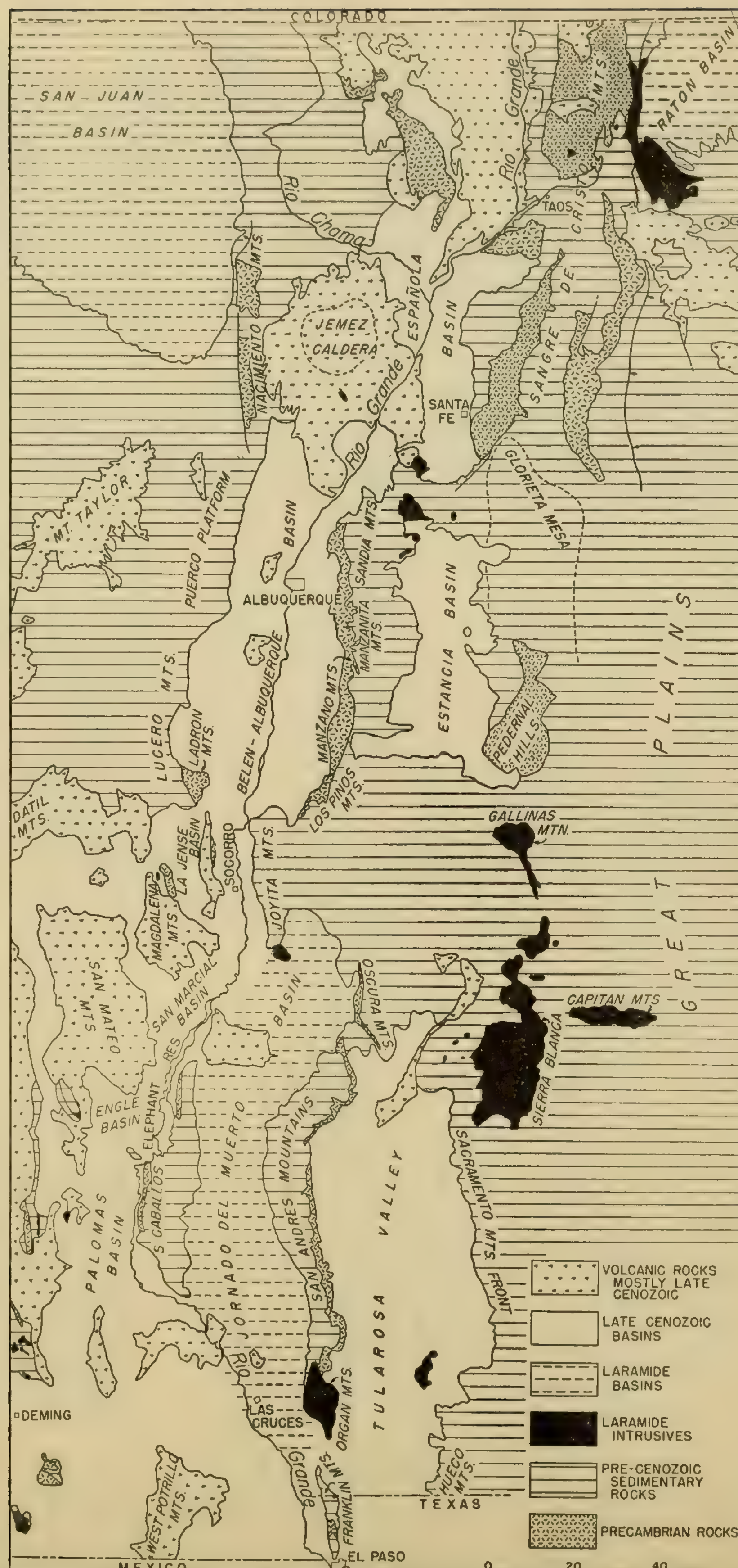


Fig. 25.11. Index map of central New Mexico.

of depressions between bordering ranges and plateaus or platforms. The basins are known collectively as the Rio Grande depression. See Fig. 25.11 for names and locations of the principal physiographic features. The ranges are the result of Laramide deformation first, and later graben faulting which has resulted in the general depression. This north-south belt is generally recognized as the southward continuation of the Rockies from Colorado, as well as a graben or rift belt of late Cenozoic age. The Colorado Plateau lies on the west and the Great Plains on the east.

The basins of the Rio Grande depression have an approximate *en echelon* arrangement, and the principal ones are as follows beginning on the north in southern Colorado: 1, San Luis; 2, Española; 3, Belen-Albuquerque; 4, San Marcial; 5, Engle; and 6, Palomas. The northern end of the Belen-Albuquerque basin is called the Santo Domino basin. The Estancia basin is separated from the Glorieta Mesa by normal faults, and although shallow and irregularly alluviated, it is probably part of the rift belt. The Jornado del Muerto basin, however, is not part of the rift belt, but mainly a Laramide downwarp between two Laramide uplifts. The Tularosa Valley is a downfaulted basin, and although not part of the Rio Grande depression, is associated with it tectonically, and is part of the general rift belt.

Major Laramide Structures

If the map of Fig. 25.11, and particularly the *Geologic Map of New Mexico* (1928) are studied, a major anticlinal uplift is suggested by the geology of the Sacramento Mountains front, the Oscura Mountains, and the San Andres Mountains. The Tularosa Valley, which is mostly a downfaulted basin, appears to have formed essentially in the core of the large uplift. The uplift evidently had folds within it because several small islands of Pennsylvanian strata appear in the alluvium well out in the basin, and therefore, one cannot assume that erosion had stripped the central part of the broad uplift everywhere to the Precambrian before the graben faulting occurred. See cross sections S and T of Fig. 25.16. The postulated uplift is labeled the San Andres in Fig. 25.12.

It has been assumed by some geologists that the westward tilted strata

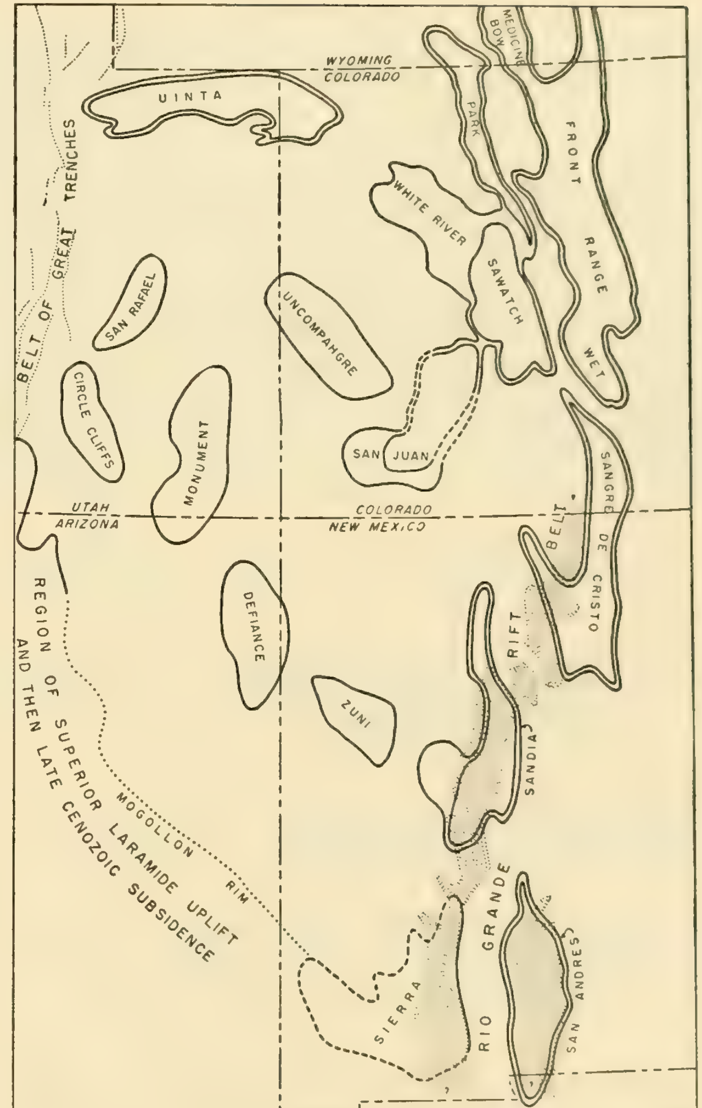


Fig. 25.12. Laramide uplifts and late Cenozoic belts of rifting around the Colorado Plateau. The uplifts with extensively exposed Precambrian cores are shown by double line; those with Paleozoic core principally, by single line. Refer to Fig. 25.8 for New Mexico, and Fig. 25.2 for Colorado. Rift valleys stippled.

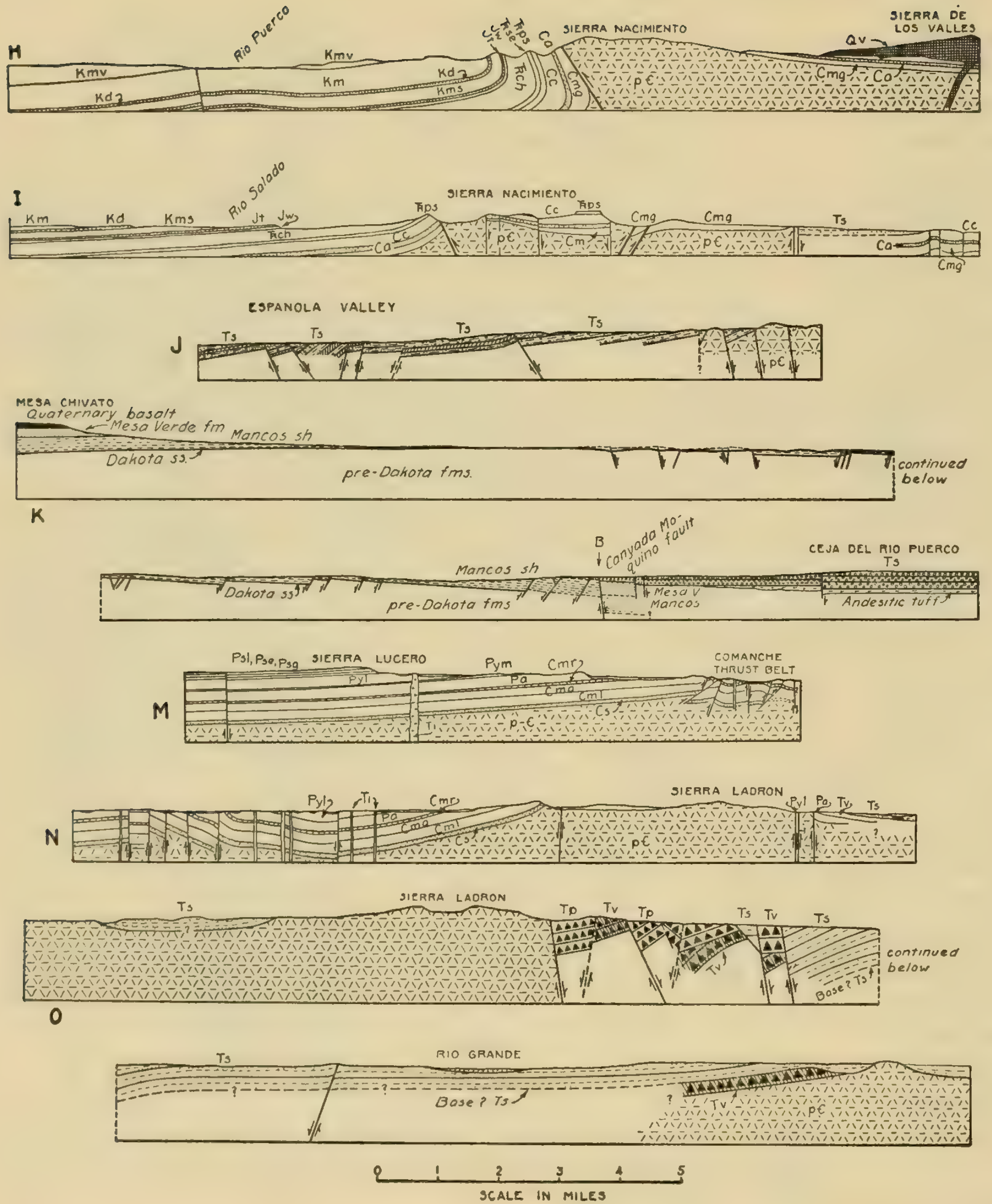


Fig. 25.13. H, section across the northern Sierra Nacimiento. After Renick, 1931.

I, section across the southern Sierra Nacimiento. After Renick, 1931. Cmg, Magdalena group; Ca, Abo sandstone; Cc, Chupadera formation; Ꞧch, Chinle formation; Jw, Wingate sandstone; Jt, Todilto formation; Kms, Morrison formation; Kd, Dakota sandstone; Km, Mancos shale; Kmv, Mesa Verde formation; Qv, Qauternary volcanoes of the Sierra de los Valles.

J, section across Espanola Valley. After Denny, 1940b. Ts, Santa Fe formation.

K, section from Mesa Chivato to Ceia del Rio Puerco. B, Boundary between Colorado Plateau and Basin and Range provinces. After Bryan and McCann, 1938.

M, section across the Lucero uplift. After Kelley and Wood, 1946. Cs, Sandra formation; Cml, Gray Mesa member; Cma, Atrasado member; Cmr, Red Tanks member of the Madera limestone; Pa, Abo formation; Pym, Meseta Blanca member; Pyl, Los Vallos member of Yeso formation; Psg, Pse, Psl, members of San Andres formation. C is Pennsylvanian, P is Permian. Ti, Tertiary intrusives; Tv, Tertiary extrusives; Tv, Tertiary agglomerate; Ts, Miocene Santa Fe formation.

N, section through Sierra Ladron and the southern end of the Sucero uplift, here a structural basin. After Kelley and Wood, 1946. Cs, Sandra formation; Cml, Gray Mesa member; Cma, Atrasado member; Cmr, Red Tanks member of the Madera Limestone; Pa, Abo formation; Pym, Meseta Blanca member; Pyl, Los Vallos member of Yeso formation, P is Permian. Ti, Tertiary intrusives; Tv, Tertiary agglomerate; Ts, Miocene Santa Fe formation.

O, section of the San Acacia area. Tv, Tertiary volcanic flows and tuffs; Tp, Popotosa formation; Ts, Santa Fe formation. After Denny, 1940a.

in the San Andres Mountains and the eastward tilted strata in the Sacramento front are due to the block faulting, and hence that the ranges are entirely late Cenozoic or Basin and Range in age. If late Cenozoic sediments are found to rest on Mesozoic beds on the floor of the graben, then this is the proper interpretation, but if the basin fill rests on Paleozoic and Precambrian beds, which seems to be the case, then the uplift is much older than the faulting, and would be considered Laramide.

Another fairly evident large uplift is indicated by the geology of the Ladron and Lucero Mountains, Puerco Platform, and Nacimiento Mountains on the west, and the Los Pinos, Manzano, Manzanita, and Sandia Mountains on the east. Precambrian rock appears to have been extensively exposed in the core before graben faulting of the Belen-Albuquerque basin. The uplift is called the Sandia, in Fig. 25.12 after the imposing Sandia Mountains.

A thrust fault of Laramide age runs along the west side of the Nacimiento mountains, and it has resulted in the Precambrian crystallines resting on the Cretaceous (Renick, 1931; Wood and Northrop, 1946). Examine cross sections H and I of Fig. 25.13. The thrust is of fairly high angle along most of its length, but at the north end it has several imbricate slices that dip at low angles. The maximum stratigraphic throw is 3500 feet (Wright, 1946).

A thrust that flanks the east side of the south end of the Sierra Nacimiento is shown by Renick (1931) but not by Wood and Northrop (1946). See section I, Fig. 25.13. The elevated block between the two opposing thrusts contains Paleozoic strata that are somewhat folded and faulted. Northward, the general structure of the range is an asymmetrical faulted uplift, the west flank being composed of steeply upturned and overthrust beds.

A broad, faulted monocline with downthrow on the east (see Fig. 25.14) leads southward from the Sierra Nacimiento about 45 miles to the Lucero uplift, where again thrusting has been recorded. The thrust on the east front of the Lucero uplift extends from the Ladron Mountains northward 30 miles to Carrizo Arroyo, where it dies out in a tear fault (Kelley and

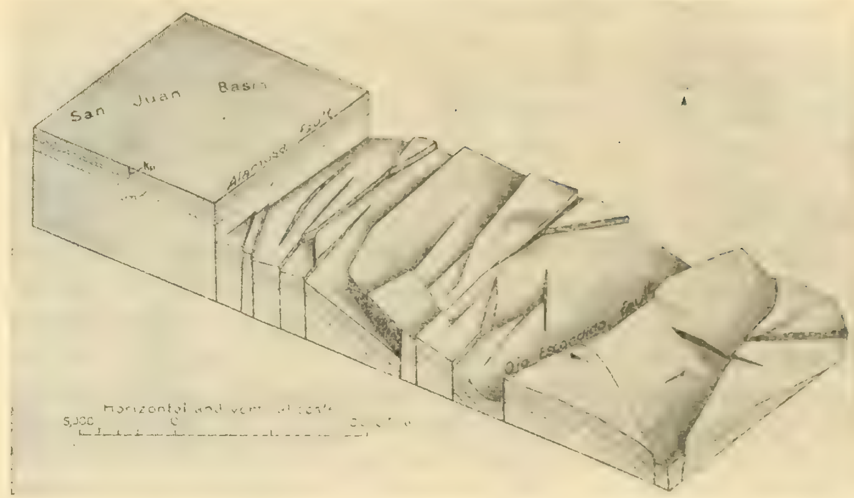


Fig. 25.14. Faulting of the monocline between the Sierra Nacimiento and the Sierra Lucero. After Hunt, 1938. Ku, Cretaceous shale and standstone; Jm, Jurassic standstone.

Wood, 1946). The thrust dips westward at angles ranging from high to low with the west side, the Lucero uplift, overthrust eastward. The stratigraphic displacement ranges from 1000 to 4000 feet.

The connecting monocline and its faults may have come into existence later than the Nacimiento and Lucero thrusts (Wright, 1946), but definitely before Miocene time (pre-Santa Fe formation, Upper Miocene).

Disconformities between the Ojo Alamo sandstone and the Torrejon formation, and between the Torrejon and the Wasatch, point to the beginnings of uplift in latest Cretaceous and early Paleocene time (Reeside, 1924). Probably the main uplift and thrusting occurred at the beginning of Eocene time, preceding the deposition of the Eocene Wasatch, and continued for some time during its deposition.

From the above it is evident that the postulated large Sandia uplift was not a simple anticline, but that it had small thrust structures within it, such as the Lucero, and possibly folds. Also the development probably proceeded in phases from latest Cretaceous into the Tertiary.

The Sangre de Cristo Range in New Mexico is less complicated than in Colorado. The eastern front resembles that of the Front Range of Colorado (Smith and Ray, 1941), and the flat-lying sedimentary formations of the plains are abruptly upturned along the mountain front, and the Dakota sandstone makes prominent ridges. See section E of Fig. 25.15. A normal fault follows the contact between the sedimentary strata and the Precambrian core of the range for more than 7 miles.

Still farther south, the structure becomes a broad arch out of which Glorieta Mesa is now eroded. See sections F and G of Fig. 25.15. The east flank is fairly sharply flexed at the north end, but toward the south through Cuervo Butte the arch is broad and regular. The low Pedernal Range, once one of the Ancestral Rockies, is covered in the Glorieta Mesa area; the overlapping of the Pennsylvanian strata on its north end is pictured in the lowest cross section. The faults along the west side of the Glorieta Mesa are younger than the arch and flexures, and are classed as Basin and Range. They probably resulted in the valley fill of the Estancia basin.

The extent of the original Sangre de Cristo uplift in Laramide times is difficult to decipher because of the graben faulting and the extensive volcanism. The rendition of it in Fig. 25.12 is very approximate.

Still another Laramide uplift, the Sierra, appears to have formed in southwestern New Mexico. Throughout its extent chiefly Precambrian and Cambrian rocks are exposed, but through it the major Rio Grande depression now exists. On the west it probably became part of the extensive Laramide uplift southwest of the Mogollon Rim, but the region is so extensively covered with Tertiary volcanics that the relations cannot be well established.

The uplifts of the Colorado Plateau and the central Laramide belt of Colorado and New Mexico as depicted in Fig 25.12 all seem to be related tectonically, and their origin by vertical uplift is emphasized. Most geologists who have mapped in the Laramide belt of Colorado and New Mexico have considered the thrust faults to indicate compressional orogeny, and especially intense compression in Colorado. The Williams Range, Gore Range, and Never Summer Range thrusts in Colorado are probably the most impressive, but these have in no respect the stratigraphic throw of the thrust sheets of the central Rockies.

Vertical uplift of the magnitude of 2 or 3 miles is indicated by the Front Range, and where the vertical movement has been abrupt along one flank or the other, and steep fronts of imposing elevation have formed, the mechanical elements for major gravity slide blocks are set up. The uplifts of the Colorado Plateau where the Precambrian rocks are not extensively exposed represent a less amount of vertical movement, and it is noted that thrusts have not formed on their margins. It seems logical to the writer, therefore, to regard the thrusts of the Laramide Rockies of Colorado and New Mexico as gravity slide phenomena.

In Chapter 33 on the igneous provinces of the western part of the continent the theory is advanced that the uplift of the Colorado Plateau and other adjacent areas in the Rocky Mountains in Laramide time and afterward was due to expansion of a column of the mantle underneath, and that this expansion was at least partly due to its partially melting. Also considerable magma made its way up to the crystalline complex, and there spread out in megasills to elevate the crust above as great blisters. These are the uplifts shown in Fig. 25.12. The concept is illustrated in Fig. 36.4. Lagging somewhat after the intrusion of the megasills came the near-surface and surficial igneous activity, so widespread throughout the Plateau and marginal areas.

Rio Grande Rift Belt

Kelley (1952) has reviewed the Rio Grande Rift Belt very well, and summarizes his conclusions as follows:

> In about middle Tertiary time volcanic activity that extruded rhyolitic to andesitic rocks developed on an enormous scale. These eruptions, together with their great outwash of alluvial material, accumulated to thicknesses of several thousand feet. The volcanic suites occur mostly in the western half of the Rocky Mountain belt and in the adjacent Colorado Plateau; but locally, as in the Raton, Cerrillos-South Mountain, and Sierra Blanca areas, the eruptions developed along the Great Plains border. Nevertheless, the uplifts bordering the east side of the depression are notably lacking in this suite of rocks. Little or no sharp folding or overthrusting accompanied the volcanic episode. High-angle faulting, however, appears to have accompanied and followed the great igneous activity. In several places there appears to have been two or three distinct volcanic stages separated by intervals of tectonic disturbance and erosion. Although local basins of accumulation appear to have developed during this epoch of Tertiary deposition and deformation, the areas of accumu-

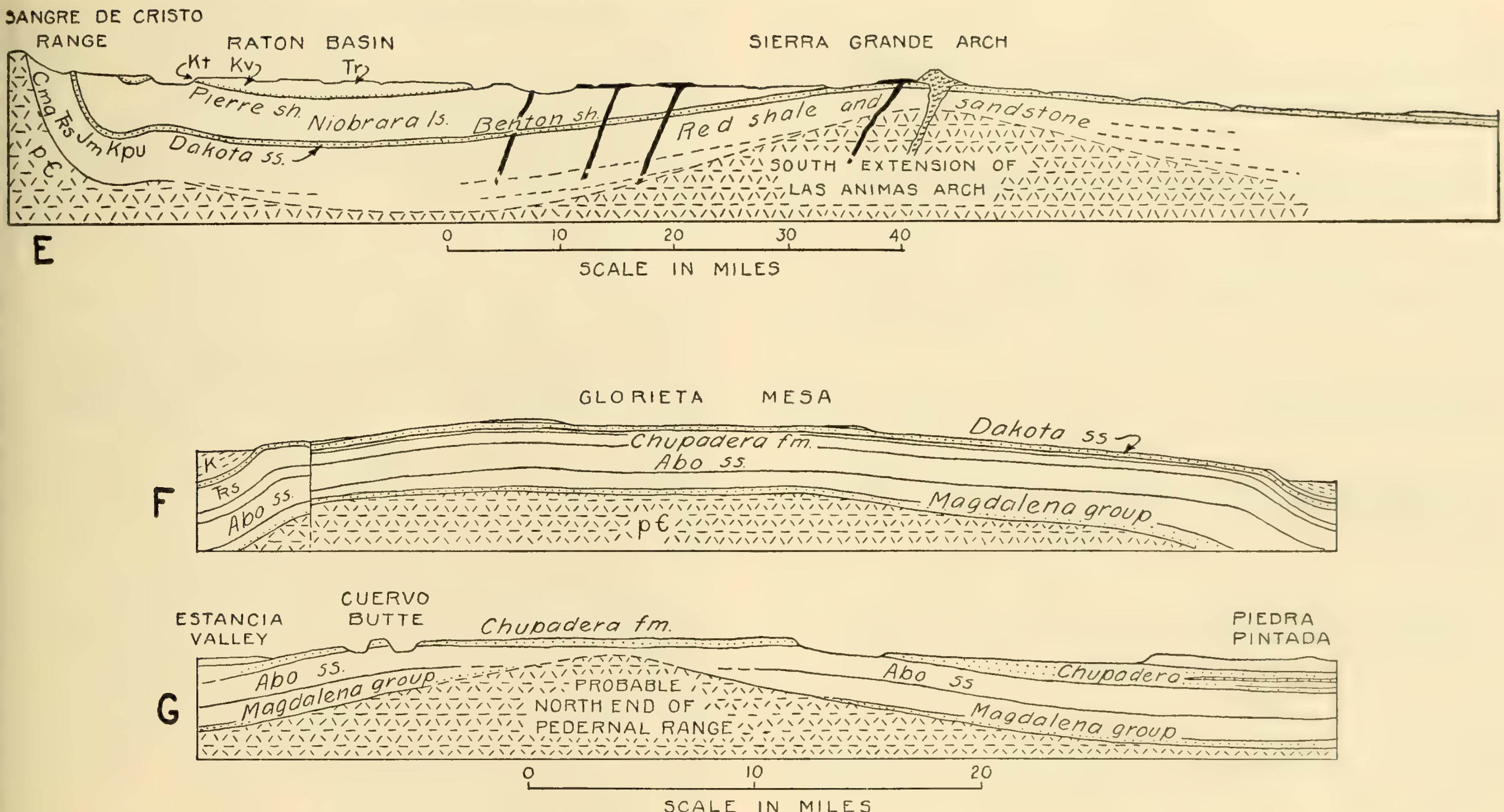


Fig. 25.15. Upper section from Sangre de Cristo Range eastward across the Raton basin and Sierra Grande arch. Modified after Darton, 1928. Cmg, Magdalena group; TRs, red shale and sandstone; Jm, Morrison formation; Kpr, Purgatoire formation; Kt, Trinidad sandstone; Kv, Vermejo formation; Tr, Raton formation.

Lower sections across the north and south ends of Glorieta Mesa, south of the Sangre de Cristo Range. TR, Triassic red shale and Jurassic, Wingate, Kayenta, and Morrison. Modified after Darton, 1928. The fault on the west side of Glorieta Mesa is probably Basin and Range in age, and younger than the Laramide.

lation appear to have been rather wide, and the troughlike aspects of the later Rio Grande depression were not yet developed. In wide areas, the middle Tertiary flows and pyroclastic and volcanic alluvial beds lie with only slight unconformity or discordance upon the earlier non-volcanic sediments. The intense fracture belt and prominent tilted blocks which are so characteristic of the Rio Grande depression and adjoining uplifts are later features.

The development of the Rio Grande structural belt probably began in late Miocene time and culminated . . . toward the end of Pliocene time. With the development of the linked *en echelon* basins the Santa Fe sediments, which are the characterizing feature of the Rio Grande depression began to

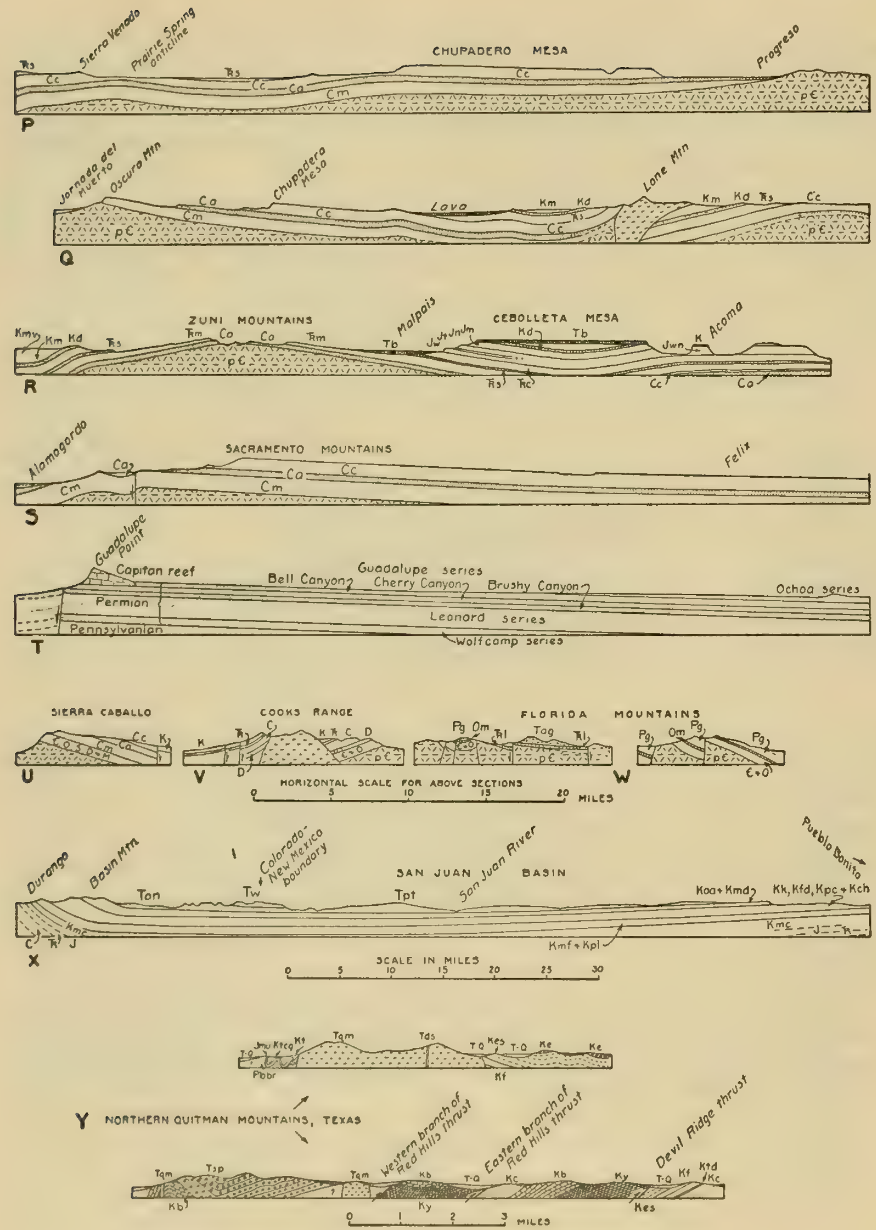


Fig. 25.16. Cross sections of south-central New Mexico and the trans-Pecos of Texas. See index map, Fig. 25.1.

P, section across Chupadera Mesa. After Darton, 1928. Cm, Magdalena group; Ca, Abo sandstone; Cc, Chupadera formation; Ŧs, Dockum group; Kd, Dakota (?) sandstone; Km, Mancos (?) shale.

Q, section across Sierra Blanca system. After Darton, 1928. Cm, Magdalena group; Ca, Abo sandstone; Cc, Chupadera formation; Ŧs, Dockum group; Kd, Dakota (?) sandstone; Km, Mancos (?) shale.

R, section across the Zuni Mountains and the Cebolleta Mesa. Ca, Abo sandstone; Cc, Chupadera formation; Ŧm, Moencopi formation; Ŧs, Shinaru conglomerate; Ŧs, Chinle shale; Jw, Wingate sandstone; Jt, Todilto sandstone; Jn, Navajo sandstone; Km, Mancos shale; Kmv, Mesa Verde formation; Tb, basalt. After Darton, 1928.

S, section across the Sacramento Mountains. Cm, Magdalena group; Ca, Abo sandstone; Cc, Chupadera formation. After Darton, 1928.

T, section from Guadalupe Point (El Capitan) eastward across Culbertson County, Tex. The Delaware basin lies east of the section. After King, 1942a.

U, section across the Sierra Caballo. After Darton, 1928. Є, Bliss sandstone; O, El Paso limestone and Montoya limestone; S, Fusselman limestone; D, Percha shale; M, Lake Valley (?) limestone; Cm, Magdalena group; Ca, Abo sandstone; Cc, Chupadera formation; K, Cretaceous beds.

V, section across Cooks Range. After Darton, 1928. Є and O, Bliss sandstone, El Paso Montoya, and Fusselman limestone; D, Percha shale; C, Lake Valley limestone, Magdalena formation; Pg, Gym limestone; Ŧ, Lobo formation; K, Sartan sandstone.

W, sections across the Florida Mountains. After Darton, 1928. Є and O, Bliss sandstone and El Paso; Om, Montoya, and Fusselman limestones; Pg, Gym limestone; Ŧl, Lobo formation; Tag, Tertiary agglomerate.

X, north-south section across the San Juan basin. After Darton, 1928. C, Pennsylvanian and Permian; Ŧ, Triassic; J, Jurassic; Kmc, Mancos shale; Kpl, Point Lookout sandstone; Kmf, Menefee formation; Kch, Cliff House sandstone; Kle, Lewis shale; Kpc, Pictured Cliffs sandstone; Kfd, Fruitland formation; Kk, Kirtland shale containing Farmington sandstone member, Kkf; Kmd, McDermot formation; Koa, Ojo Alamo sandstone; Tan, Animas formation; Tpt, Torrejon and Puerco formations, Paleocene; Tw, Wasatch formation, Eocene.

Y, sections across the northern Quitman Mountains, Texas. After Huffington, 1943. Pbbr, Briggs formation (Permian); Jmu, Malone formation (Upper Jurassic); the following formations are all Lower Cretaceous; Kt, Torcer formation; Ky, Yucca formation; Kb, Bluff formation; Kc, Cox formation; Kf, Finlay formation; Kes, Espy formation; Kc, Etholen formation; T–Q, Tertiary and Quaternary alluvium.

form. The Santa Fe has been assigned to ages that range from late Miocene to Pleistocene. In its typical development it is an alluvial-fan deposit of a characteristic pinkish or light-tan color. Although it is locally grayish, it generally stands in fairly marked contrast to the somber brown, purplish-brown, or grayish-white of the middle Tertiary sediments upon which it often rests. The Santa Fe is typically a relatively non-volcanic sediment, but in many places, especially along the west side of the depression, its coarse fragments may be almost exclusively volcanic, but even in these places the characteristic pinkish color is evident in the clay and sand beds. The Santa Fe in large part reflects the rocks which were at the surface in the adjoining uplifts, and the superposition of its local members commonly roughtly reflects, in reverse order, the stratigraphic superposition of the adjoining areas. In many places where the adjoining uplift consisted of carbonate rocks such as the Magdalena, San Andres, or lower Paleozoic formations, the adjacent Santa Fe is largely a calcirudite fanglomerate. Elsewhere playa and lake deposits form a large part of the Santa Fe. Pyroclastic breccia and tuff may be abundant in the Santa Fe, and this is especially true around the Jemez uplift. Basaltic flows are almost a characteristic of the Santa Fe, and are intercalated sparingly throughout the section.

CENTRAL NEW MEXICO PORPHYRY BELT

A zone of Laramide intrusions extends from Black Mountain in the Sangre de Cristo Range southward through central New Mexico into Mexico. It coincides with the belt of faulting and uplifts in the central part of New Mexico. The intrusions take the form of stocks, laccoliths, dikes, and sheets. The largest stocks are in the Sandia and Ortiz Mountains just southwest of Santa Fe, and in the Sierra Blanca, Capitan, and Gallinas Mountains, just north of the Sacramento uplift. Also, the Organ Mountains northwest of El Paso are mostly of intrusive rock. These plutons are clearly intrusive into the Paleozoic strata, and in places into the Upper Cretaceous, and all are believed to be Laramide. Like the uplifts, most of them have not been accurately dated because the youngest rocks intruded are commonly Paleozoic. See Chapter 36 for further discussion of the igneous rocks.

GUADALUPE AND MARATHON UPLIFTS

An arm of the Sacramento uplift extends southeastward to the Texas boundary, where the Guadalupe Mountains compose themselves and extend south-southeastward about 80 miles to the Davis Mountains volcanic area. The Guadalupe Mountains separate the salt basin on the west from the Delaware basin on the east. See cross sections S and T of Fig. 25.16. Like the other uplifts, they are an asymmetrical structure, an eastward-tilted block with a complex zone of high-angle faults along the west side. Examine the *Tectonic Map of the United States.*

The Davis Mountain volcanic area is in effect a large structural basin. Along its northeast border is a belt of long, fairly gentle anticlines and synclines that lead to the Marathon uplift, in whose core the late Paleozoic compressional structures known as the Marathon Mountains are exposed.

The Marathon uplift is dome-shaped and surrounded by Cretaceous beds. The exposed core is not centrally located in the Cretaceous dome; it is mostly in the western half. According to King (1937), the base of the Lower Cretaceous strata 4 miles north and south is 5500 feet below the base in the Glass Mountains that form the north rim. The dip of the Cretaceous beds to the east is about 100 feet per mile.

The western margin of the Marathon dome is formed by the Del Norte and Santiago mountains, which are eroded out of a sharp monocline or anticline. The fold is overturned toward the west and broken in most places by an eastward-dipping thrust that has raised Paleozoic and Lower Cretaceous rocks on the east against younger Cretaceous beds on the west (King, 1937). This narrow belt of compressional deformation seems to be alone in the belt of domes and basins of central New Mexico and the Trans-Pecos region of Texas. About 75 miles to the southeast, the Sierra Madre Oriental of Coahuila, Mexico, which is a system of folded and thrust structures, continues in the strike of the Santiago thrust, and may be a continuation of it. The Sierra Madre Oriental, however, seems more related to the Laramide folding and thrust belt of southern Arizona, southwestern New Mexico, and the Quitman Mountains region of Texas-Mexican border, southeast of El Paso and northwest of the Davis Mountains volcanic area. See the map of the Laramide orogenic belts, Fig. 19.1.

The Marathon dome is younger than the Lower and Upper Cretaceous beds, but only part of the elevatory movement preceded the effusion of the Davis Mountain volcanics, which are Eocene and Oligocene in age. In one place, the lavas overlie Upper Cretaceous strata, and in another

they rest on the Lower Cretaceous. Also, the lavas do not dip away from the dome as steeply as the underlying Cretaceous beds, but since the lavas *have* been tilted, part of the doming followed their outpouring. Furthermore, King (1937) points out, an anticline that has folded the Permian and Cretaceous rocks in the Glass Mountains extends northwestward into the Davis Mountains, where it involves the Tertiary lavas. The folds are definitely older than the normal faults that break the strata in the Glass, Del Norte, and Santiago Mountains, and there offset the lavas. The normal faulting appears to have taken place soon after the lavas were poured out, possibly in Oligocene or Miocene time.

26.

COLORADO PLATEAU

GENERAL GEOLOGY

The Colorado Plateau is one of the world's show places, not only for the tourist but for the geologist. Badlands, high escarpments, and deep gorges leave few geologic secrets covered if they are searched for. Early geologists such as Gilbert, Powell, and Dutton, who first explored the Plateau geologically, made it classical territory. Their line drawings of the physiographic features and their diagrams of the structures still stand as masterpieces. The contributions in the years since these early investigations have been on the details.

A bird's-eye view of part of the province may be obtained from a stereogram of Gilbert's (1877), reproduced here in Fig. 26.2. Pictures of the Grand Canyon, Zion National Park, the natural bridges of San Juan County, Cedar Breaks, and Bryce National Monument are commonplace and serve to identify small spectacular parts of the great region. The major structures of the plateau are shown in the index map of Fig. 26.1.

The paleotectonic and paleogeologic maps of this book show clearly that the Colorado Plateau was a shelf area adjacent to the westward lying Cordilleran geosyncline. Parts of Arizona were Precambrian terrane until the Pennsylvanian. From Mississippian times to the close of the Cretaceous, shallow seas covered the entire province, with the exception of the Uncompahgre and Zuni ranges of the Ancestral Rockies, which were finally buried in Triassic time.

The Paleozoic section of the Grand Canyon is given in Fig. 26.4, and a number of Mesozoic and late Paleozoic sections of various parts of the Colorado Plateau are reproduced in Figs. 26.5 and 26.6. References to detailed stratigraphic studies in the plateau may be obtained from the two figures.

The paleogeographic and tectonic development of the region in Paleozoic time has been treated in Chapters 6 and 15.

In brief, the pre-Laramide history of the Plateau is as follows. When the Cambrian seas invaded the area, a relief in places, at least, of about 800 feet existed (Sharp, 1940), and the surface was entirely buried by the Cambrian sediments. In general, the absence of Ordovician and Silurian strata throughout the Plateau, with only a disconformity marking their place, indicates either gently emergent conditions during these periods or that, toward the close of the Silurian, the region became emergent, and any beds that were deposited during the interval were removed. The Mississippian was one of limestone deposition, but the Pennsylvanian was one of considerable crustal unrest with the building of the Ancestral Rockies and the subsidence of the Paradox basin. The western margin of the Plateau was the transition from shelf to miogeosyncline in Paleozoic time and later in Cretaceous and early Tertiary time the site of mountain building and accumulation of thick orogenic deposits. See Chapter 22 on the Central Rockies.

Although horizontal strata dominate the landscape, several monoclines were formed which are the steep flanks of large asymmetrical anticlines, some 30 miles across and 100 miles or more long.

Fig. 26.1. Major geologic features of Colorado Plateau. Black areas are lava fields; stippled areas are Early Tertiary sediments; horizontally dashed areas are Cretaceous sediments; and cross-ruled areas are laccolithic mountains. S.R.S., San Rafael Swell; C.C.U., Circle Cliffs uplift; S.L.C., Salt Lake City; P., Provo; F., Filmore; B., Beaver; Fl., Flagstaff; PC., Prescott. Lava fields in the Great Basin not shown, especially in the St. George area.

Another significant type of structure in the Colorado Plateau is the laccolithic mountain. There are several clusters of laccolithic intrusions, and these have produced the mountains known as the Henry, La Sal, Abajo (or Blue), La Plata, Ute, and Carrizo. See index map, Fig. 26.1 and Fig. 33.7. Also, another high mountain, Navajo, is probably a laccolithic structure. There are several major volcanic fields, one in the High Plateaus of Utah; one just south of the Grand Canyon in Arizona, the San Francisco Mountains; and one in eastern Arizona and western New Mexico, the Datil field. The flexures, laccoliths, volcanic fields, and other features are described in the following pages.

ASYMMETRICAL ARCHES AND BASINS

In a very broad way the Monument uplift, with Permian beds extensively exposed in the core, is the center of the Plateau, and is nearly surrounded by Cretaceous and Tertiary basins. Auxiliary uplifts break the continuity of the surrounding basins or render the general pattern irregular (Figs. 26.1 and 19.2). The San Rafael Swell, the Circle Cliffs uplift, and the Uncompahgre uplift lie just inside the Cretaceous and Tertiary basins on the west, north, and northeast, and the Kaibab (Fig. 26.8) and Defiance (Fig. 26.9) uplifts break the continuity of the Cretaceous basins on the south and southwest. The uplifts in Utah and the Kaibab in Arizona are characterized by a sharp monoclinal flexure on the east side, broad tops, and gently dipping west flanks (Fig. 26.7). In the high, desert climate of the Plateau these monoclines are grand features of the scenery. The arches are all believed to be of Laramide age, and are crossed indiscriminately by the master streams that drain the Plateau. The streams are, therefore, either superposed or antecedent. The greatest scenic spectacle in the Plateau, in the minds of many people, is the gorge of the Colorado River (Grand Canyon) across the Kaibab uplift. The river here has cut through the entire Paleozoic section and also well into the Precambrian.

The Cretaceous and Tertiary basins, aside from the Black Mesa of Arizona, are bounded on the outside by major Laramide uplifts, and the deepest parts or troughs of the basins lie close to these uplifted and mountainous areas. The Tertiary section in the Uinta Basin is 10,000

Fig. 26.2. Waterpocket monocline, Henry Mountains, and parts of the Kaiparowits and Aquarius plateaus and the Colorado River, all in Utah. Reproduced from Gilbert, 1877.

Fig. 26.3. Waterpocket monocline and Henry Mountains as if no erosion had occurred since the folding and intrusions. The same area is depicted structurally in this illustration as physiographically in the opposite illustration. Reproduced from Gilbert, 1877.

Era	Period and Epoch	Group and Formation	Kind of Rock	Prevailing Color	Thickness in Feet
Mesozoic	Triassic	*Shinarump congl.*	*Conglomerate*	*Brown*	*25*
Mesozoic	Triassic	*Moenkopi fm*	*Shale and sandstone*	*Red*	*480*
		UNCONFORMITY			
Paleozoic	Carboniferous: Permian	*Kaibab limestone*	*Limestone and sandstone*	*Gray, buff, and red*	*525*
Paleozoic	Carboniferous: Permian	*Coconino sandstone*	*Sandstone*	*Light buff*	*350*
Paleozoic	Carboniferous: Permian	*Hermit shale*	*Sandy shale*	*Red*	*225*
Paleozoic	Carboniferous: Permian	*Supai formation*	*Sandstone and shale with some limestone*	*Red and gray*	*825*
		UNCONFORMITY			
Paleozoic	Carboniferous: Miss.	*Redwall limestone*	*Limestone*	*Gray stained red*	*450 to 500*
Paleozoic	Devonian	*Temple Butte ls.*	*Limestone and sandstone*	*Pale purplish*	*0-36*
		GREAT UNCONFORMITY			
Paleozoic	Cambrian (Tonto group)	*Muav limestone*	*Sandy shale and limestone*	*Gray*	*100*
Paleozoic	Cambrian (Tonto group)	*Bright Angel shale*	*Sandy shale*	*Greenish gray*	*450 to 650*
Paleozoic	Cambrian (Tonto group)	*Tapeats sandstone*	*Sandstone*	*Brown*	*225*
		GREAT UNCONFORMITY			
Proterozoic	Algonkian	*Unkar and Chuar groups of Grand Canyon series*	*Sandstone and shale with some limestone; contains sheets and dikes of lava*	*Mostly red*	*0 to 12000*
		GREAT UNCONFORMITY			
Proterozoic	Archean	*Vishnu schist*	*Schist, granite, and gneiss*	*Dark gray*	*Not known*

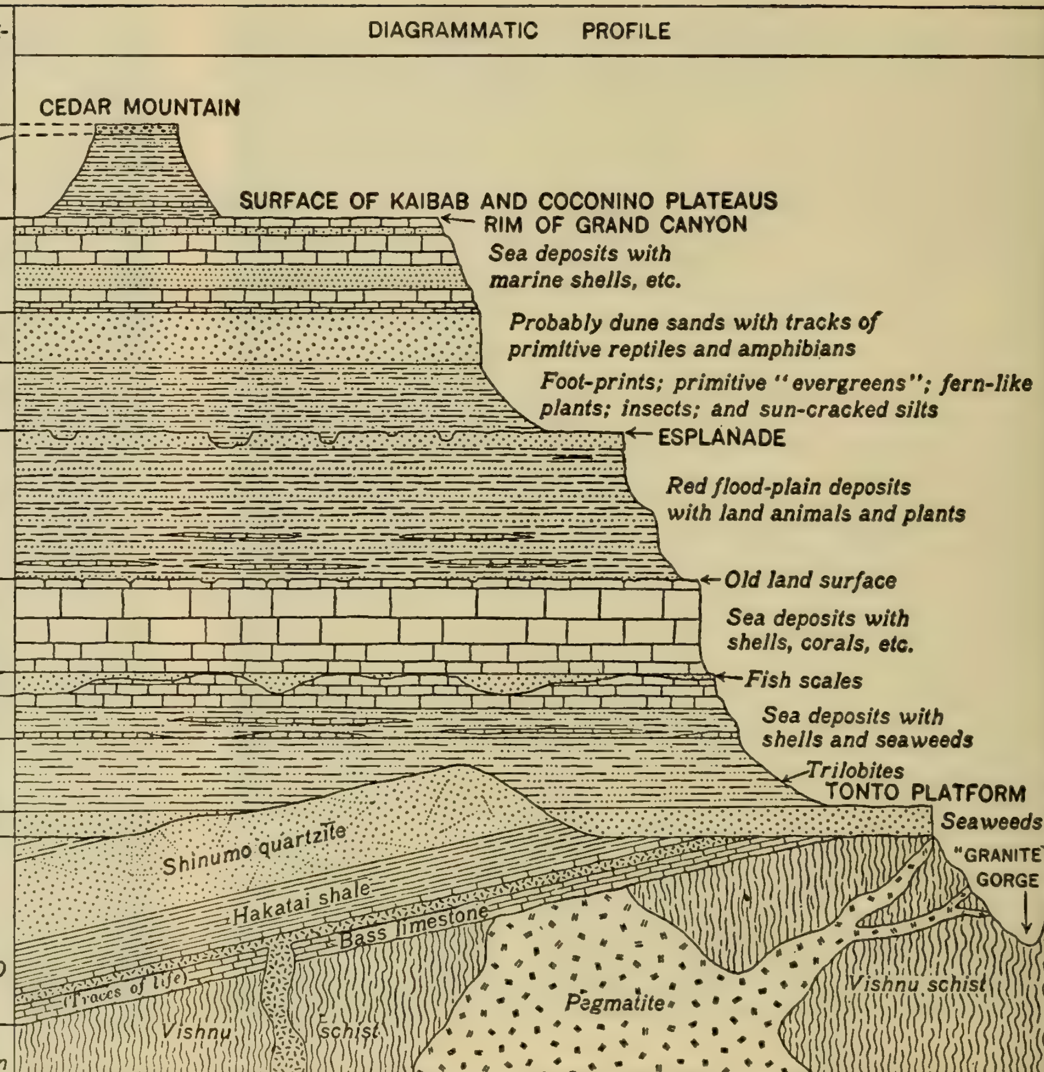


Fig. 26.4. Generalized columnar section of rocks forming the walls of the Grand Canyon of the Colorado. After Noble, 1924.

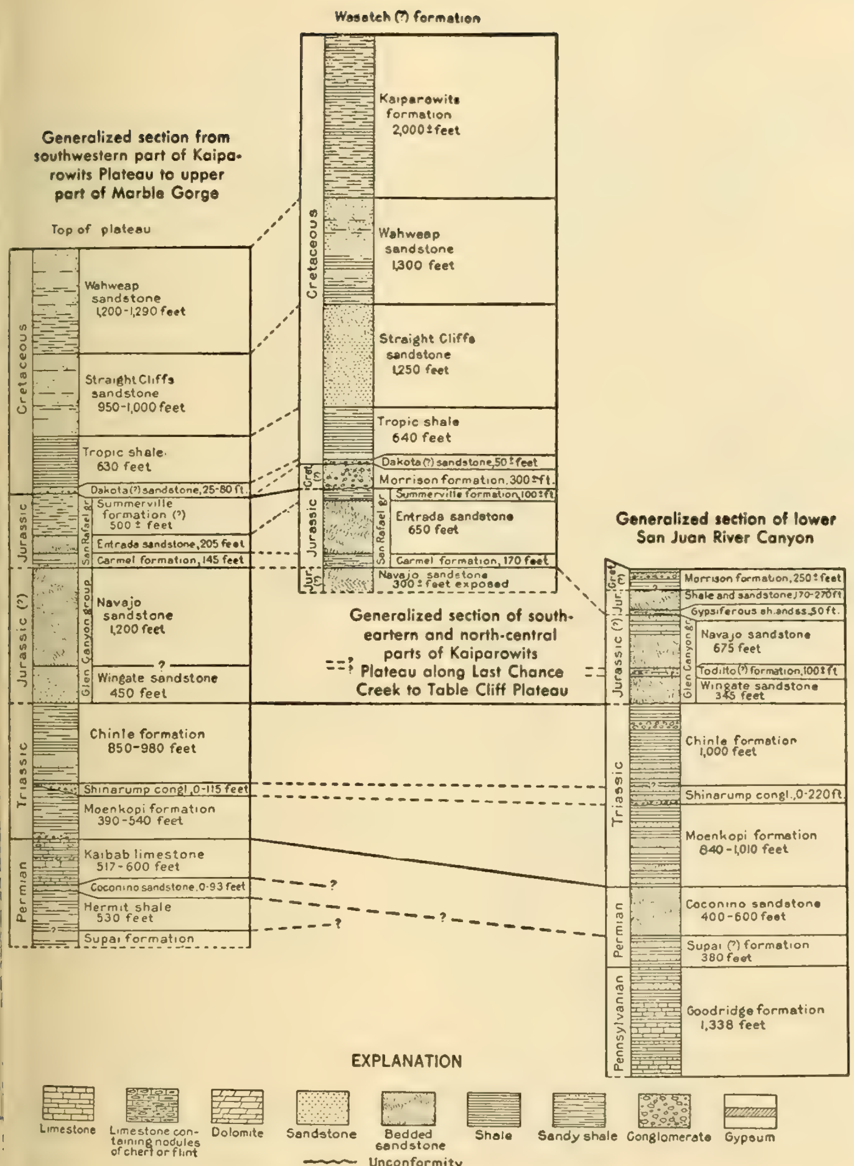


Fig. 26.5. Paleozoic and Mesozoic formations of the Kaiparowits Plateau and lower San Juan River Canyon. After Gregory and Moore, 1931.

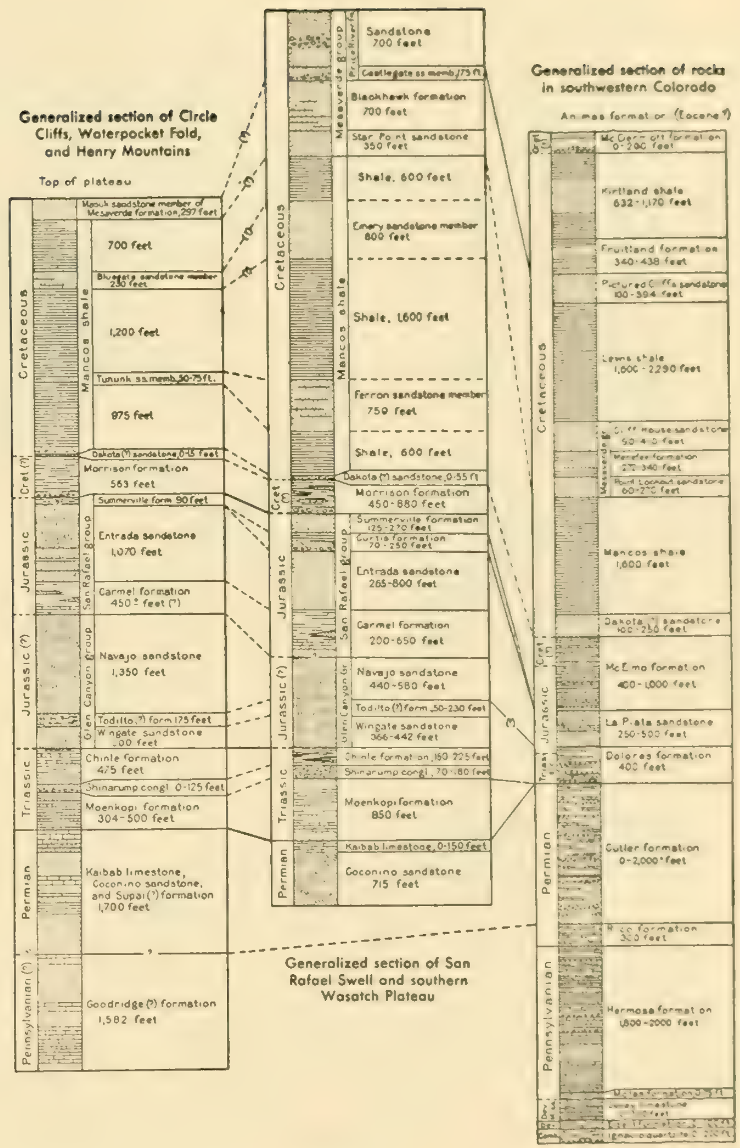


Fig. 26.6. Paleozoic and Mesozoic formations of Circle Cliffs uplift, Henry Mountains, San Rafael swell, Wasatch plateau, and southwestern Colorado. After Gregory and Moore, 1931.

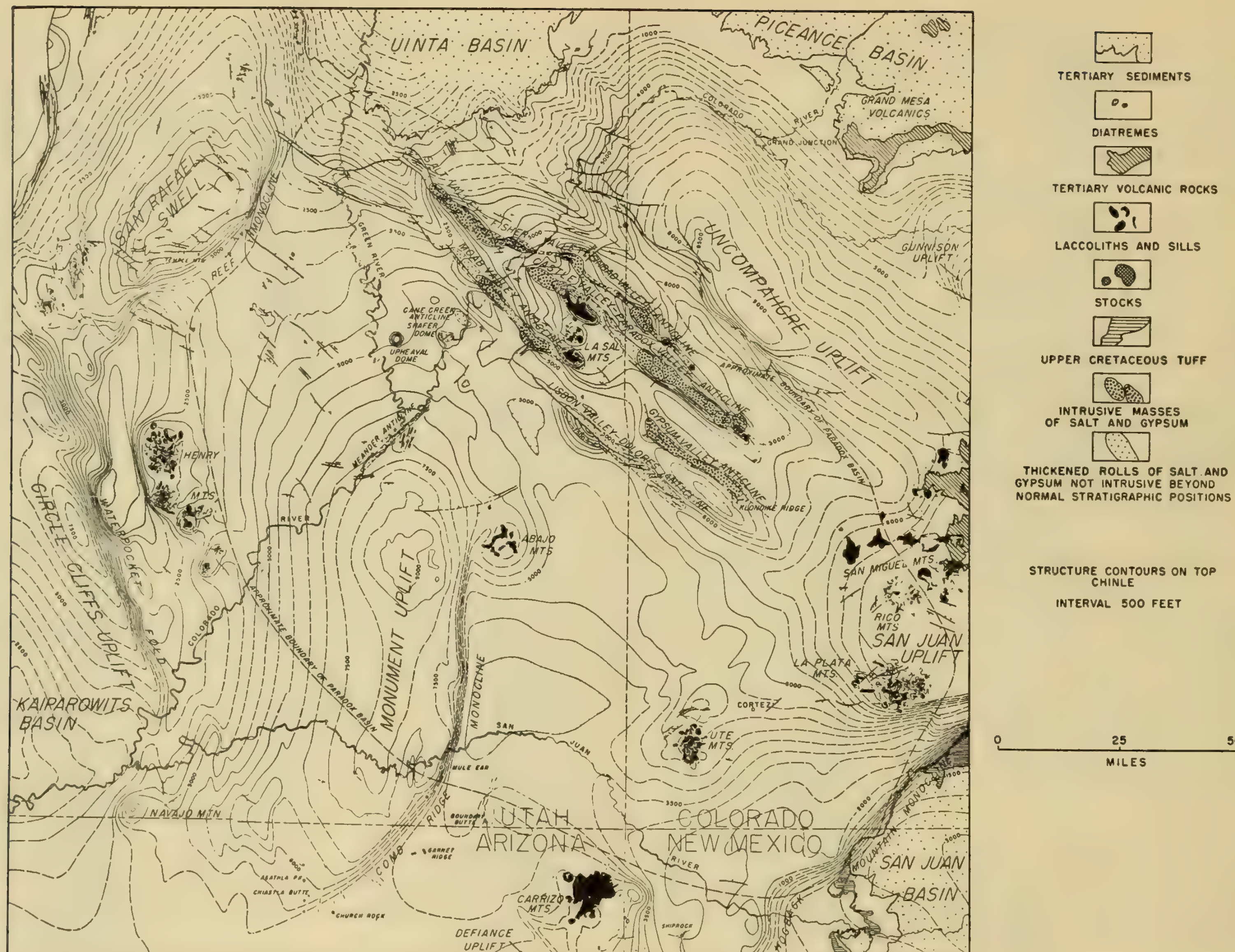


Fig. 26.7. Tectonic map of central part of Colorado Plateau. Compiled by Shoemaker, 1954.

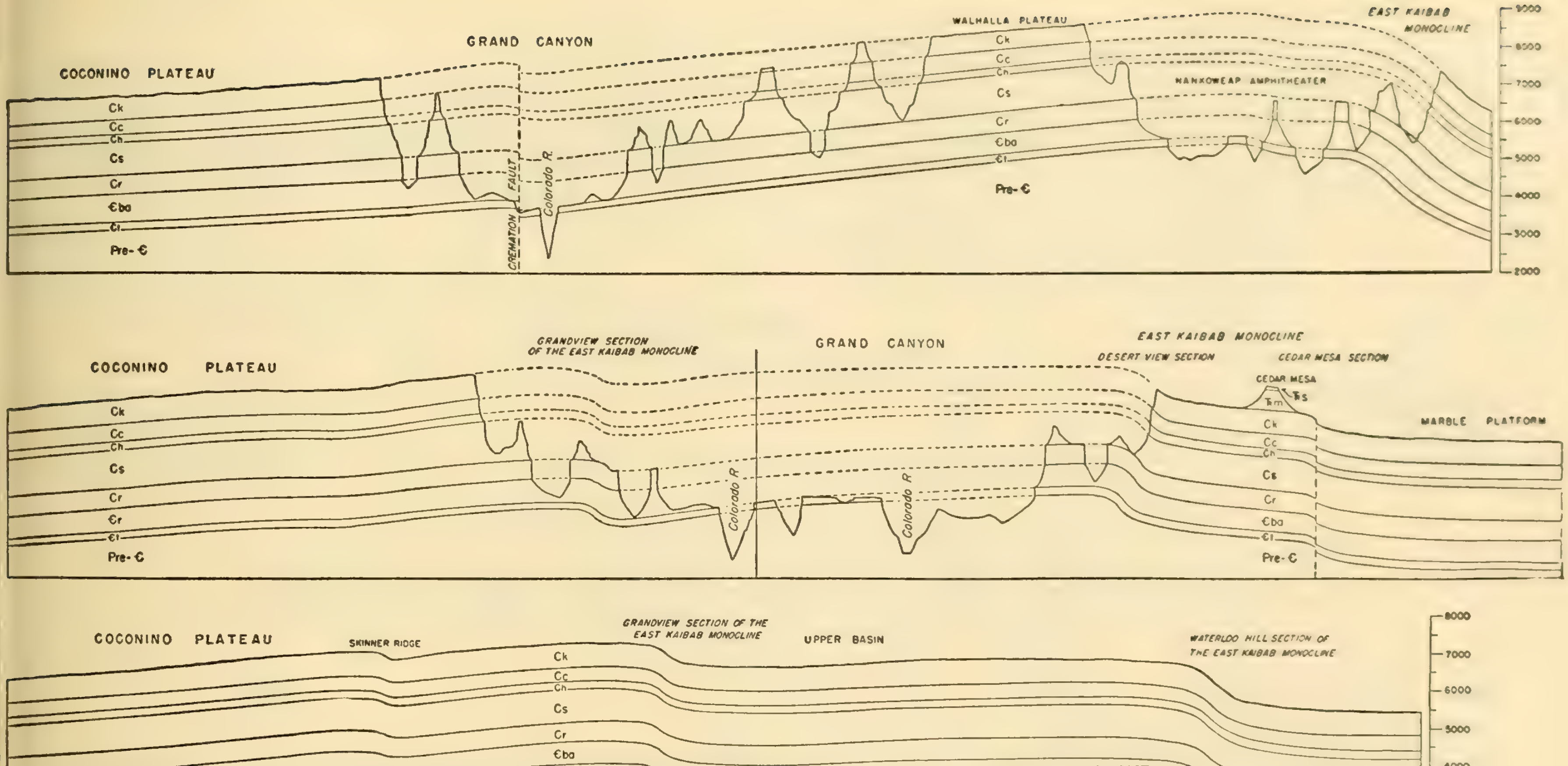
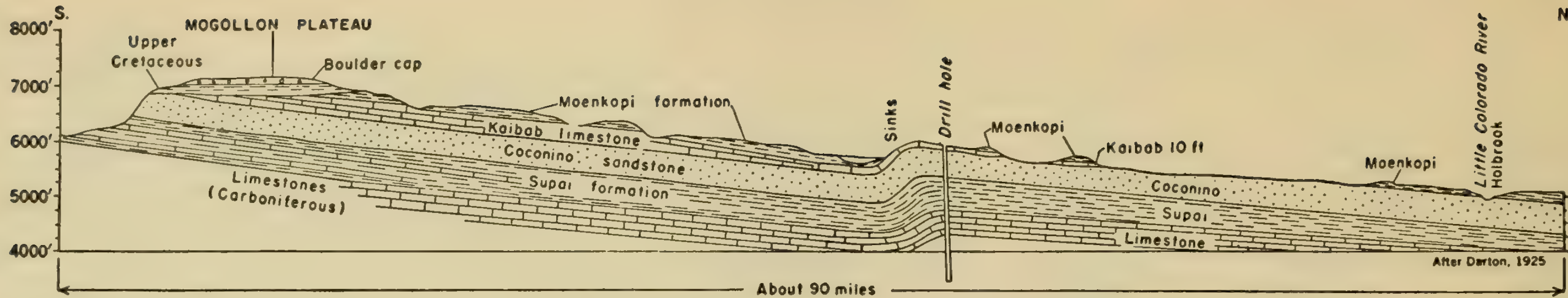


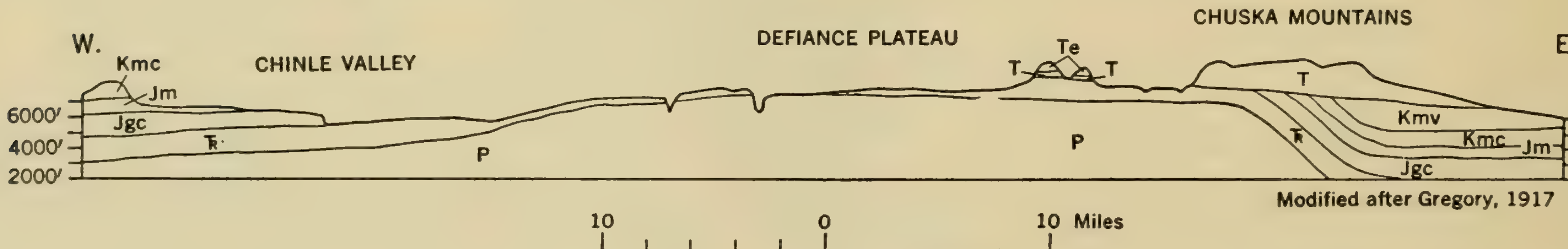
Fig. 26.8. Cross sections from north to south of the Kaibab uplift and East Kaibab monocline. After Babenroth and Strahler, 1945. Cr, Redwall ls.; Єba, Bright Angel sh.; Єt, Tapeats ss.

feet thick (Fig. 26.10). A great Eocene fresh-water lake was impounded by the crustal movements around the Uinta arch, and in it the petroliferous strata of the Green River formations were depoisted. The Upper Cretaceous and Lower Tertiary strata of the High Plateaus are very thick (Chapter 22), and the Cretaceous of the San Juan basin is about 10,000 feet in maximum thickness (Fig. 26.11).

The major monoclines displace the strata vertically 5000 to 8000 feet. They die out gradually at the ends, and curve in toward the uplift. These features have been taken to mean that the Precambrian basement was upfaulted and that the flexible sedimentary beds were draped over the faults (Baker, 1935). In places, the faults break through to the surface. Small faults are numerous in the uplifts and basins.

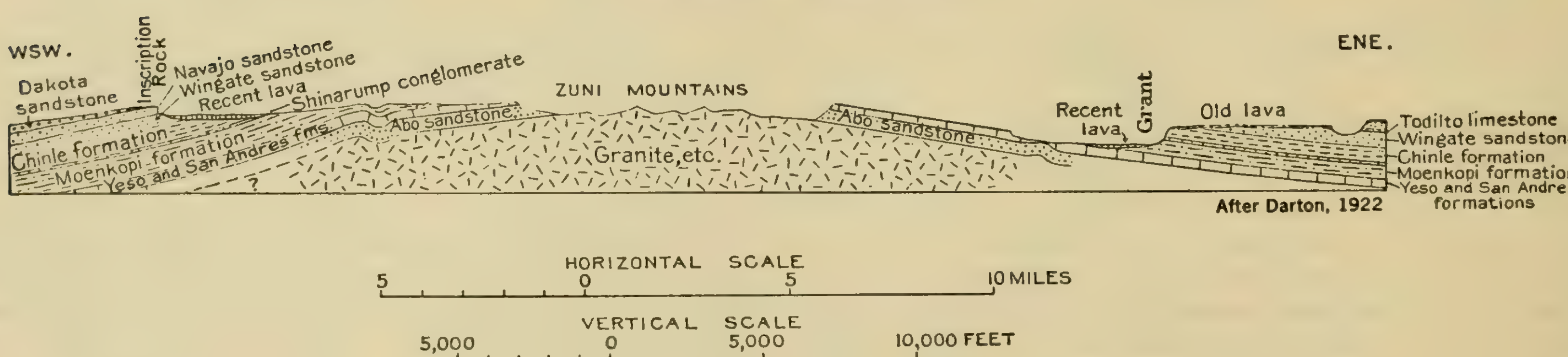


Cross section of Mogollon Rim, illustrating overlap of Upper Cretaceous formations southward onto Permian formations.



Te, late Tertiary lavas; T, Chuska sandstone; Kmv, Mesaverde group; Kmc, Mancos shale; Jm, Morrison formation and San Rafael group; Jgc, Glen Canyon group; Ꞧ, Triassic formations; P, Permian formations

Cross section of the Defiance uplifts.



Section across the Zuni Mountain upwarp, New Mexico.

Fig. 26.9. Some uplifts of the Colorado Plateau. Reproduced from Hunt, 1956. Chuska ss. is Pliocene in age.

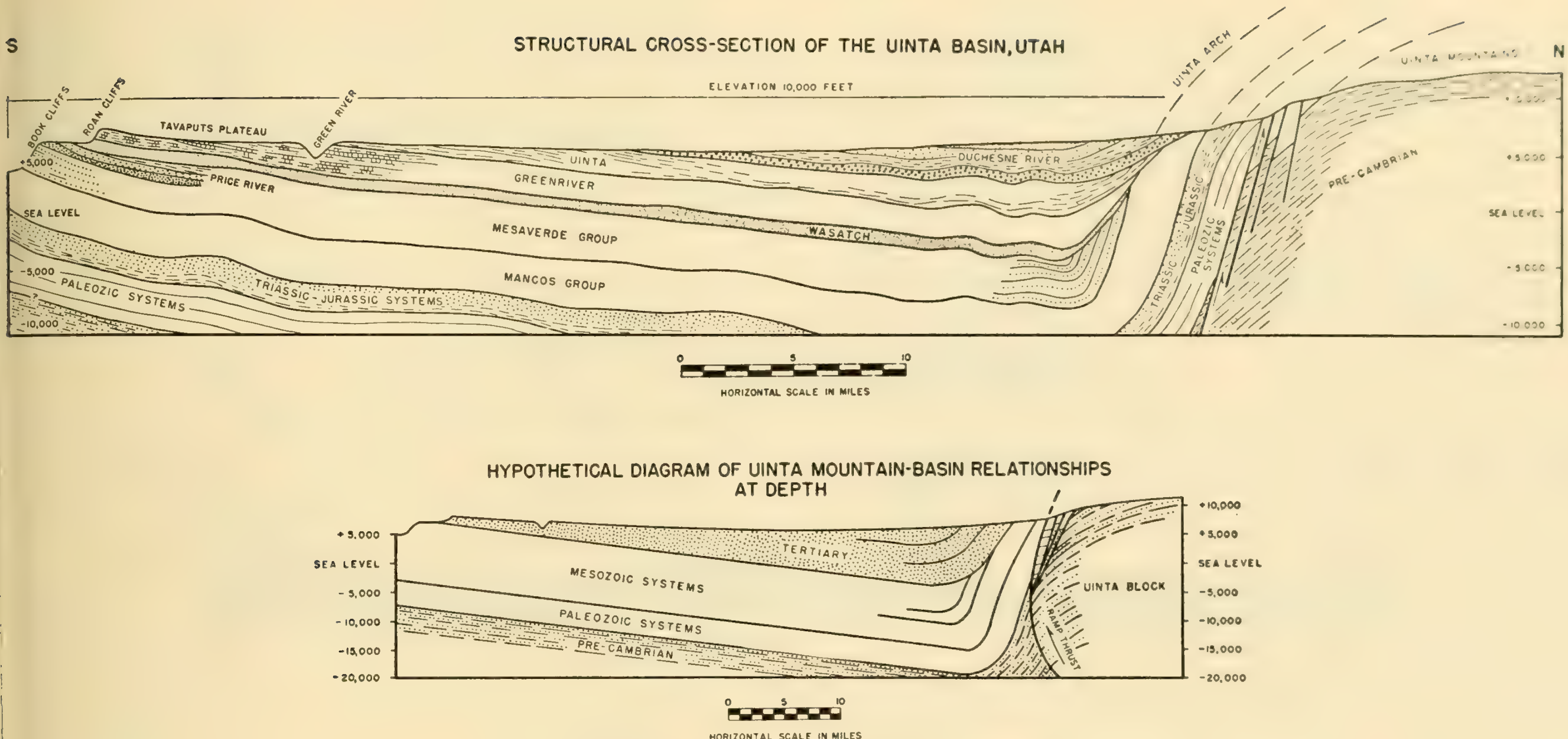
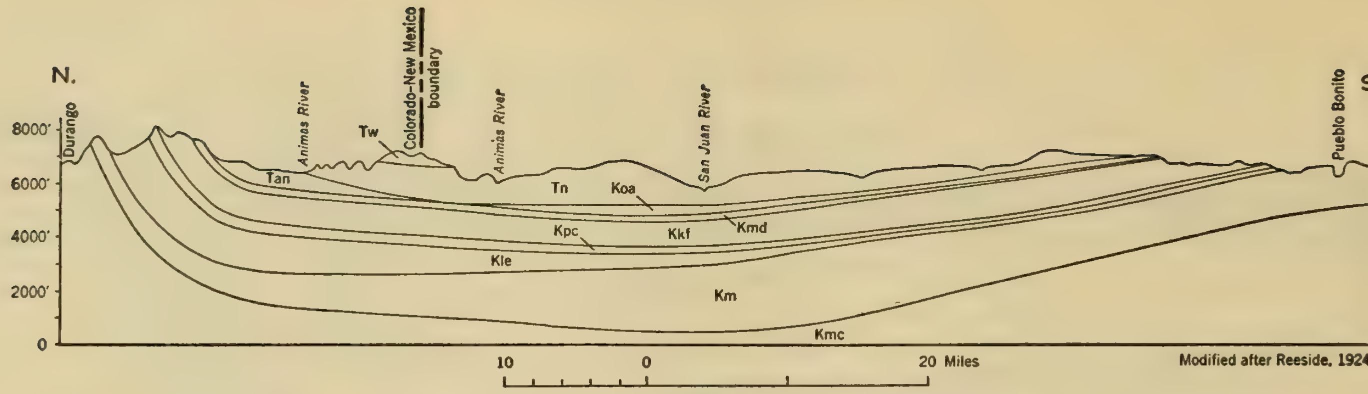


Fig. 26.10. Section of the Uinta basin approximately through Vernal and southward along the Green River. Compiled by Orlo Childs and P. T. Walton.

SALT ANTICLINES

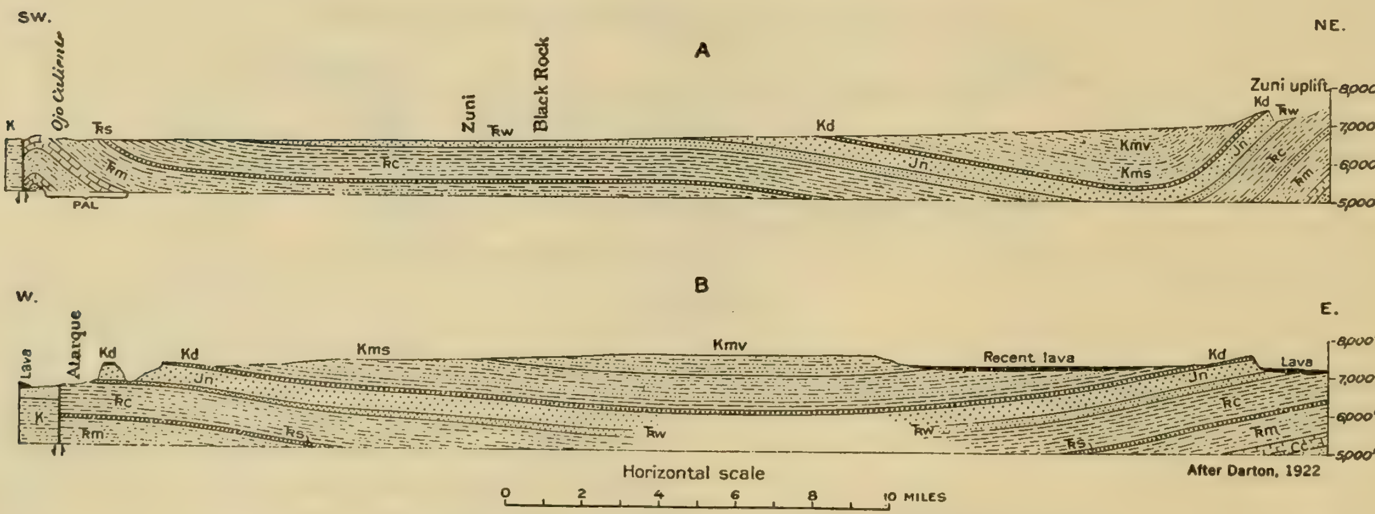
A zone of flexures and faults extends in a northwest-southeast direction from east-central Utah into west-central Colorado. They consist basically of eight elongate anticlines variously modified by collapse folds and normal faults. Their position and orientation are shown in Figs. 26.7 and 26.12. Between the anticlines are simple and gentle synclines. Stokes (1948) observes that their strike is common with that of the Uncompahgre Range of the Ancestral Rockies and suggests that this relation points to a system of deep-seated breaks in the basement rocks. The problem is very complex, however, because the faults and folds (the Meander anticline) have a northeast direction, and the great monoclinal flexures a northerly trend.

It is fairly certain that the structures are closely connected with salt flowage and solution. At certain times and places, the salt and gypsum have acted as ordinary rock and have suffered the same deformation as



Tw, Wasatch formation; Tn, Nacimiento formation (Puerco and Torrejon faunal zones): Tan, Animas formation; Koa; Ojo Alamo sandstone; Kmd, McDermott formation; Kkf, Fruitland formation and Kirtland shale; Kpc, Pictured Cliffs sandstone; Kle, Lewis shale; Km, Mesaverde group; Kmc, Mancos shale

Diagrammatic section across the San Juan basin from Durango, Col., to Pueblo Bonito, N. Mex.



K, Cretaceous formations; Kmv, Mesaverde group; Kms, Mancos shale; Kd, Dakota sandstone; Jn, Navajo sandstone; Ꞓw, Wingate sandstone; Ꞓc, Chinle formation; Ꞓs, Shinarump conglomerate; Ꞓm, Moenkopi formation; PAL, Upper Paleozoic rocks

Sections across the Gallup-Zuni basin, New Mexico. A, near Zuni; B, near Atarque.

Fig. 26.11. Some basins of the Colorado Plateau. Reproduced from Hunt, 1956. For the San Juan basin see also Fig. 25.13.

contiguous strata, but at other times and places the saline beds have behaved independently and, through plastic flow and solution, have produced large-scale structures in the overlying beds that have no expression in the underlying strata.

Taking the Gypsum Valley anticline as an example, Stokes (1948) re-

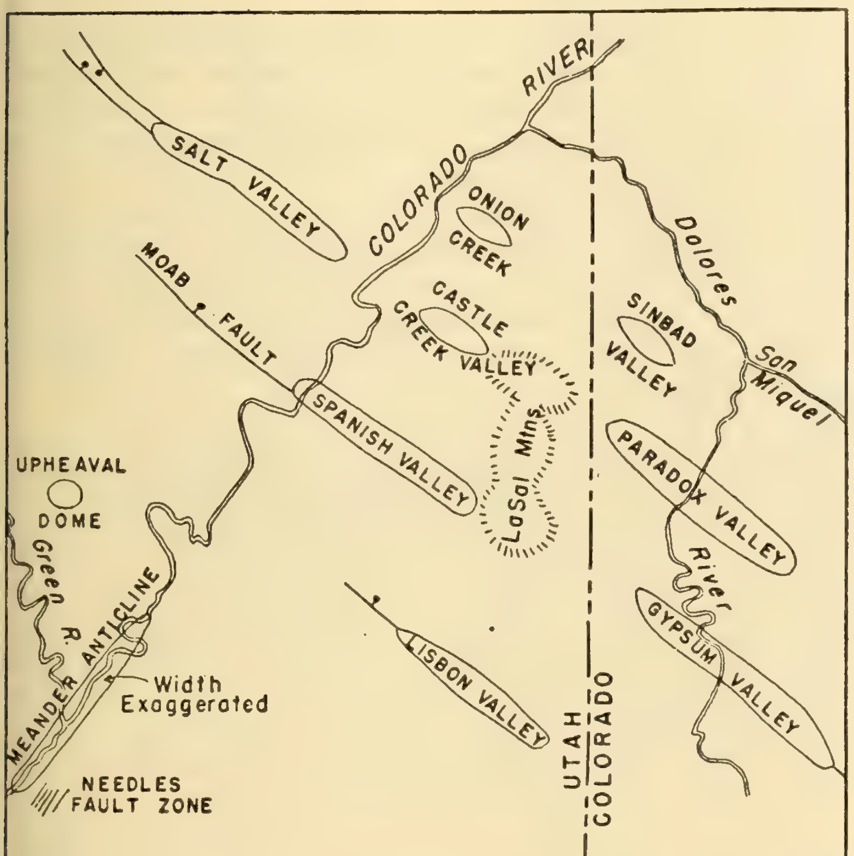


Fig. 26.12. Sketch map of eastern Utah and western Colorado showing principal structures due to salt flowage and solution. Reproduced from Stokes, 1948.

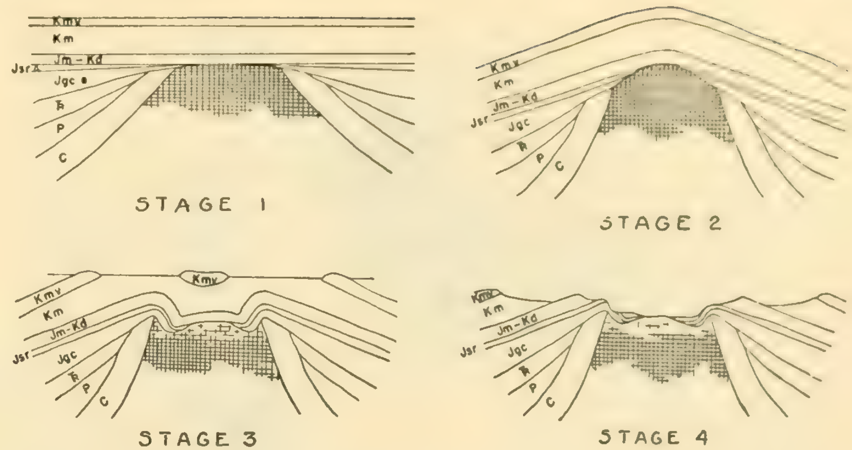


Fig. 26.13. Development of the salt anticlines in the Colorado plateau. Reproduced from Stokes, 1948. Vertical scale exaggerated. C, Hermosa formation; P, Permian formations; Ti, Triassic formations; Jgc, Glen Canyon group; Jsr, San Rafael group; Jm-Kd, Morrison to Dakota formations; Km, Mancos shale; Kmv, Mesaverde group.

cords the following development (see series of cross sections in Fig. 26.13).

1. Deposition of salt and gypsum in Paradox formation of late Pennsylvanian age. Deposition of covering Hermosa limestone beds.
2. At end of Pennsylvanian or in early Permian time, the salt pushed upward and domed the Hermosa; the Hermosa was eroded and the salt exposed.
3. The late Permian Rico and Cutler formations were deposited around the dome or anticline.
4. The salt dome was eroded nearly to a peneplain by late Triassic, and the Triassic and early Jurassic formations overlapped across the edges of older formations around the dome.
5. The late Jurassic sediments practically submerged the salt mass, and then the late Cretoceous formations were deposited undisturbed over the site of the salt mass. These five steps are all recorded in the first cross section of Fig. 26.13.

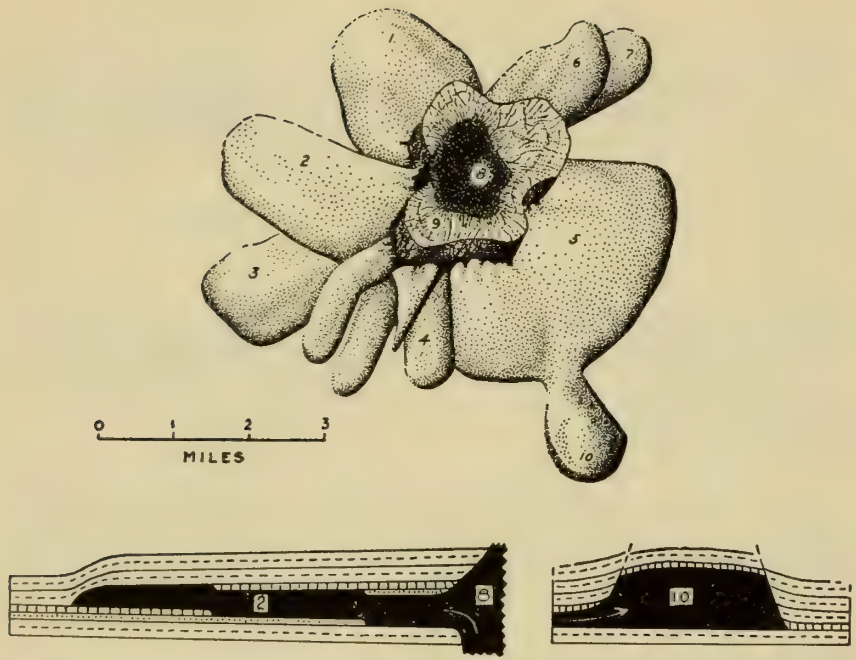


Fig. 26.14. Ground plan and cross sections of intrusive bodies of Mt. Ellen, Henry Mountains, Utah. 1, Corrall Ridge laccolith; 2, Durfey Butte laccolith; 3, South Creek Ridge laccolith; 4, Slate Creek laccolith; 5, Cooper Ridge laccolith; 6, Granite Ridges laccolith; 7, Butler Wash laccolith; 8, Mt. Ellen stock; 9, shatter zone; 10, Ragged Mountain bysmalith. Cross sections illustrate relation of a laccolith, such as No. 2, to the sedimentary rocks, and also of a bysmalith, such as No. 10. The top of the stock has been cut off to show the shatter zone and hard rock. The lower cross sections are somewhat enlarged views of the igneous bodies 2 and 10, and show the relations to the sedimentary beds. After Hunt, 1954.

6. In the early Tertiary, the entire section above the salt was arched up, and the salt intruded slightly but not more than through the late Jurassic formations.
7. The entire Colorado Plateau was elevated in mid-Tertiary time, and coincident with the uplift the laccolithic intrusions occurred which are described under the next heading. As a result of the uplift, groundwater commenced to circulate more freely, the salt masses were subjected to solution, and collapse of adjacent and overlying strata occurred.
8. With the uplift of the plateau, several partial cycles of erosion have followed, and the salt anticlines have been excavated and in places a gypsum residue over the salt exposed. See last diagram of Fig. 26.13.

LACCOLITHIC MOUNTAINS

The Henry Mountains were first described by Gilbert (1877), and with his publication they became classical for the laccolithic type of mountain. He pictured the laccolith as a tack- or mushroom-shaped, concordant pluton—as a centrally thickened sill which has arched the beds above it and has been fed through a conduit from below.

Hunt (1954) has restudied the Henry Mountains in detail and shows them to be concordant tongues extending outward from a central, trunklike stock. They are like semicircular, conical fungus growths on tree trunks. The largest cluster is illustrated in Fig. 26.14. Ninety-five percent of the intrusive rock is diorite porphyry, and the rest is monzonite porphyry. According to Hunt,

> The Henry Mountains are located in the structural basin that is one of the major folds of the Colorado Plateau [Fig. 26.15]. The basin is the antithesis of the adjoining Circle Cliffs uplift and San Rafael Swell, being of the same size and form, only inverted. The basin is sharply asymmetric and its trough is crowded against the steep west flank; the deepest part is 8,500 feet structurally lower than the neighboring uplifts.
>
> Each of the divisions of the Henry Mountains is a structural dome several miles in diameter and a few thousand feet high. In general, the domes have smooth flanks, but on most of them are superimposed many small anticlinal noses that were produced by the laccoliths. At the center of each of the domes is a stock, around which the laccoliths are clustered. The stocks are cross-cutting intrusions, mostly surrounded by a shatter zone, which consists of highly indurated sedimentary rocks irregularly intruded by innumerable dikes, sills, and irregular masses of porphyry.
>
> The dome of Mount Ellen, the largest dome in the Henry Mountains, has a broad, plateau-like upper surface that is marked with many small anticlinal folds. On this mountain the laccoliths were injected in all directions from the stock. The Hillers dome is the highest and steepest of the domes and the anticlinal folds over the laccoliths around it are restricted to the north and northeast

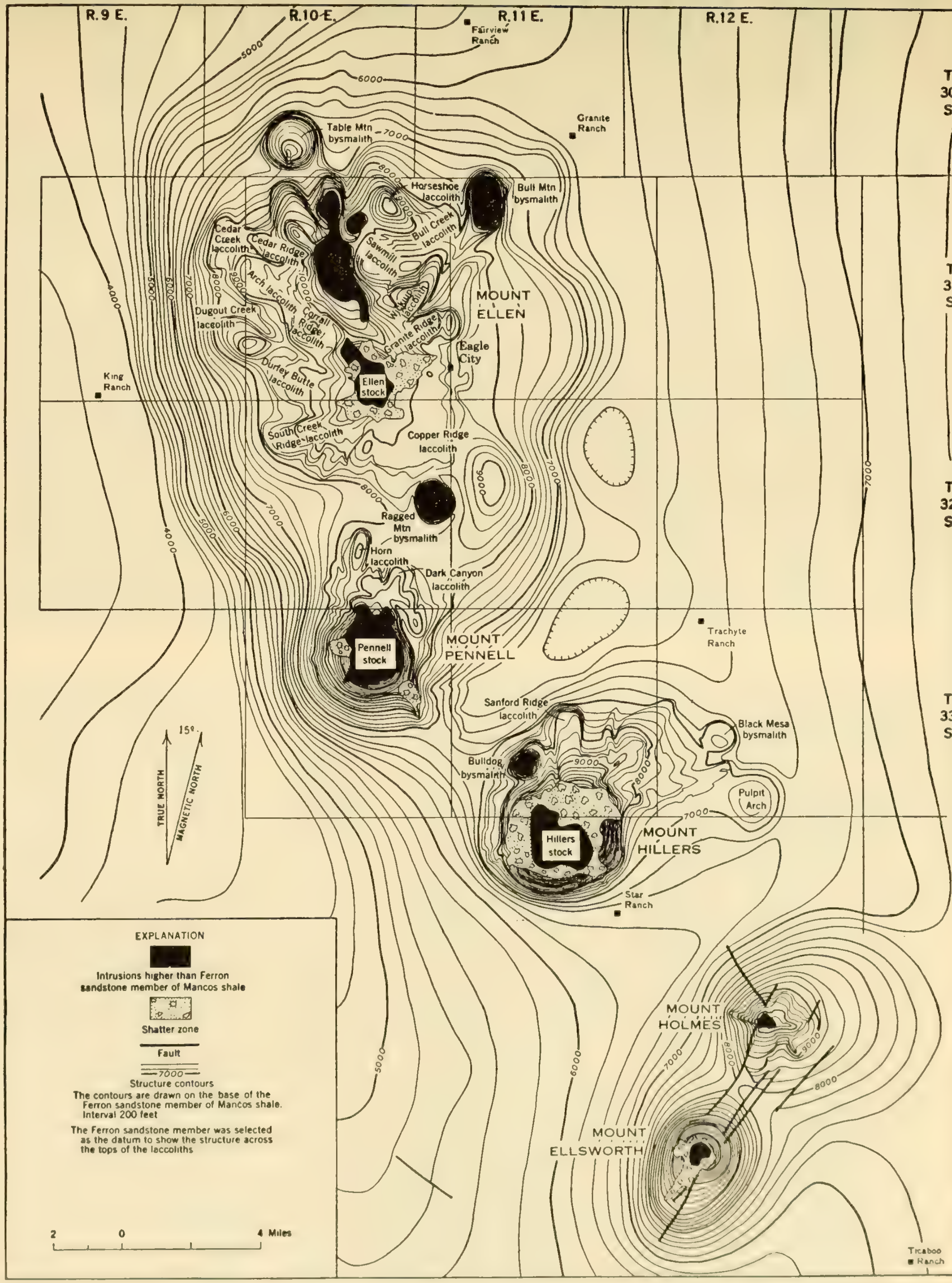


Fig. 26.15. Structure contour map of the Henry Mountains. Reproduced from Hunt, 1956.

sides. Mount Pennell is similar to Mount Hillers, only smaller. The Holmes dome is broken by faults, and its top is wrinkled with some minor anticlinal noses. The Ellsworth dome has no anticlinal folds to mar its symmetry, although it is broken by a few faults.

Several lines of evidence indicate that the laccoliths were injected radially from the stocks: (1) the laccoliths are tongue-shaped in plan and make a radial pattern around the stocks; (2) dike-like ridges on the roofs of the laccoliths trend away from the stocks; (3) the laccoliths are bulged linearly and the axes of the bulges radiate from the stocks.

Coherence and competency of the invaded rocks appears to have been an important factor controlling the stratigraphic distribution of the laccoliths. Pre-Jurassic formations, estimated to be about 5,000 feet thick, consist of well-bedded, relatively coherent, alternating thin competent and incompetent units in which very few laccoliths were injected. Overlying this is 1,200 feet of competent and highly coherent sandstone of the Glen Canyon group (Wingate, Kayenta and Navajo formations), containing even fewer laccoliths than the underlying zone. Overlying this sandstone is the San Rafael group and lower half of the Morrison formation of Jurassic age, aggregating a thickness of about 1,000 feet, and consisting of incoherent, incompetent, poorly-bedded rocks and interbedded competent layers. About 15 per cent of the total volume of the laccoliths is in these formations. The sequence including the upper half of the Morrison and the Cretaceous formations, aggregating a thickness of about 2500–3000 feet, consists largely of incoherent, incompetent shale in very thick units separated by thin competent layers. By far the greatest number of laccoliths, and at least 70 per cent by volume, are in this zone. Through this zone the laccoliths are concentrated along the thin competent layers.

Because the stratigraphy and structure of the Colorado Plateau is fairly uniform, the similarity in form of intrusion, geologic structure, and igneous rock types at the several laccolith mountains in the Plateau reflect close similarity of the igneous processes involved. The mountains are believed to represent a series of examples of one igneous process that was arrested at various stages of completion.

Navajo Mountain, a structural dome containing no exposed igneous core, represents the least advanced stage of the process. Mount Holmes, whose dome is slightly higher than Navajo Mountain and whose center contains a small stock from which a moderate number of dikes, sills, and very small laccoliths radiate, represents the next more advanced stage. The process is still farther advanced at Mount Ellsworth, where the doming is steeper and higher than at Mount Holmes and where a moderate size stock surrounded by a shatter zone occupies the center of the dome and abundant dikes and sills intrude the flanks. On Mounts Pennell and Hillers which illustrate the next most advanced stage, the doming is much steeper than on Mount Ellsworth, the stocks at the centers of the domes are much larger, the flanks of the domes contain abundant dikes and sills and in addition, to the north and northeast, huge linear, tongue-like laccoliths were injected. The dome of Mount Ellen covers a much greater area than the domes of the other mountains, and the laccoliths radiate in all directions from the stock (Hunt, 1954).

The La Sal Mountains are made up of three separate masses both topographically and geologically, and each mass consists of a stock and a cluster of radiating concordant intrusions into the sediments (laccoliths). According to Hunt (1958):

> The intrusions are in the midst of a series of salt anticlines and synclines whose axes trend northwest. Although the folding and attendant faulting in the area around the La Sal Mountains are chiefly the result of late Late Cretaceous or early Tertiary deformation, the structural history is complicated because there has been repeated plastic deformation of the salt beds and the strata arched over them. These structures antedate the intrusions and are not believed to be causally related to them.
>
> North La Sal Mountain is located on an anticline, South La Sal Mountain is in a faulted syncline, and Middle La Sal Mountain is in an area of gentle homoclinal dips between these two structures. The North La Sal Mountain forms a dome 10 miles long and 5 miles wide, and the uplift on it exceeds 6,000 feet. This dome is greatly elongated northwesterly, parallel to the axis of the anticline in which it is located. The Middle Mountain dome is nearly circular in plan, about 5 miles in diameter, and about 3,500 feet high. At Mountain dome is 6 miles long, 4 miles wide, and about 6,000 feet high. At the center of each of these domes is a stock, and radiating from each stock are laccoliths. The domes are attributed to the physical injection of the stocks. In the North and South La Sal Mountains the laccoliths spread in the salt beds of late Paleozoic age; in the Middle La Sal Mountain the laccoliths spread in shale of late Cretaceous age.

The petrology of the laccolithic groups is discussed in Chapter 33, but of interest here is the conclusion reached by Hunt that in the closing stages of fusion, crystallization, and intrusion of the North Mountain stock four pipelike masses of explosion breccias formed as diatremes were blasted through the arched roof.

The Abajo Mountains consist of two unequal parts, the smaller and northern division being an isolated dome known as Shay Mountain. It is believed to be underlain by a stock. The southern and main division consists of two parts, East Mountain and West Mountain, where central stocks are exposed with surrounding shatter zones and clusters of radiating laccoliths. A fourth stock is postulated although not exposed, and altogether around the four stocks 31 laccolithic intrusions have been

mapped (Witkind, 1958). They are mostly intrusive into the Morrison, Burro Canyon, Dakota, and Mancos formations.

UPHEAVAL DOME

Upheaval dome is a small circular structure of most peculiar and spectacular nature. It is described by McKnight (1940) as follows:

> The Upheaval dome . . . lies about 4 miles east of the Green River at the head of a short canyon (Upheaval Canyon) that cuts through the Wingate cliff (Fig. 26.7). It is circular in ground plan and consists of a conical dome surrounded by a ringlike syncline. The diameter of the dome, measured through center from the axis of the syncline on one side to the axis on the other, is 2 miles. The outer flank of the syncline is uniformly half a mile wide, making the complete diameter of the affected area 3 miles. Outside of the very sharply defined line along which the strata dip in abruptly toward the syncline, the regional low dip to the north has been undisturbed. The inward dip on the outer flank of the syncline is generally between 15° and 30°; the outward dip on the central dome ranges between 30° and 90°, though generally between 40° and 60°.

Upheaval dome is considered by McKnight to be a salt dome rather than a cryptovolcano, because of the occurrence of thick salt beds at moderate depth, and because deformation apparently took place slowly. Gravity and aeromagnetic surveys related to the geology have led Joesting and Plouff (1958) to the following conclusions:

> 1. Uplift totalling some 2,000 feet of comparatively dense, magnetic basement rock at Upheaval Dome and Grays Pasture. The uplift took place before the deposition of the White Rim sandstone member of the Cutler formation of Permian age, and may have coincided with tectonic activity during Pennsylvanian and Permian times in other parts of the Paradox basin.
> 2. Formation of a salt dome centered at the present Upheaval Dome, possibly controlled by the basement uplift. Plastic flow of salt continued, probably intermittently, until late Triassic (Wingate time).
> 3. Further doming due to salt flow, possibly in Tertiary time, coincided with renewed flow of salt in the nearby salt anticlines. The rim syncline formed during this period as a result of thinning of salt around the dome and subsidence of everlying beds.
> 4. Intrusion of igneous rock into the salt dome, probably coincident with the late Tertiary igneous activity in other parts of the Colorado Plateau (Hunt, 1956, pp. 42–53). The igneous intrusion was comparatively small. It did not displace all of the salt in the core of Upheaval Dome, but it may have been responsible for some of the additional upward movement.

VOLCANIC FIELDS

Peripheral Fields

Several volcanic fields lie around the periphery of the Colorado Plateau: the High Plateaus field in southwestern Utah, the Unikaret or Mt. Trumbull field of northeastern Arizona, the San Francisco field of north-central Arizona, the Datil field of southeastern Arizona and southwestern New Mexico, the Mt. Taylor and Jemez fields of northwestern New Mexico, the San Juan field of southwestern Colorado, and several small fields in western Colorado. See particularly Fig. 33.7. These are all described in Chapter 33.

Hopi Buttes and Navajo Volcanic Fields

Scores of volcanic necks, dikes, and lava-capped mesas rise from the high plateau of northeast Arizona and the adjacent parts of Utah and New Mexico. These are the remnants of a volcanic field that formerly covered many thousands of square miles. Erosion has so far dissected this field that the original cones have disappeared, the sheets of lava have been dismembered, and the old volcanic conduits now rise as giant towers, revealing their inner structure with singular clarity (Williams, 1936). The largest volcanic cluster is the Hopi Buttes, an isolated field 35 to 40 miles on a side (Fig. 33.7).

The surface on which the flows of the Hopi Buttes area were erupted was one of low relief and is now nowhere far from an elevation of 6000 feet. According to Williams (1936):

> Such valleys as existed on this old surface must have been choked by showers of ash during the opening stages of volcanic activity. The streams were repeatedly dammed, forming playas and ponds, which seem to have been united ultimately into a lake of great extent. This may be spoken of as Hopi Lake. Its deposits stretch as far north as Jedito Wash and south to the vicinity of the Five Buttes, a distance of almost 35 miles; in an east-west direction they are traceable for 50 miles, from near Ganado to Dilkon.
>
> Although the original cones have long since been removed, there is no difficulty in recognizing where they once existed. Erosion, acting rapidly on the surrounding sediments, has left the crater- and conduit-fillings as conspicuous towers, the summits of which cannot be more than a few hundred feet below the tops of the former cones. It is not surprising, therefore, that the intrusive rocks are indistinguishable from the surface flows.

At least 30 volcanic necks have been recognized. A few are isolated, but most occur in clusters or are aligned in a northwest-southeast direction, parallel to the adjacent dikes. The typical neck is circular in plan, though some are oval and merge gradually with dikes.

The typical Hopi vent was opened by the explosive drilling of a cylindrical pipe, and doubtless a pyroclastic cone or maar-like depression was formed about the orifice. Subsequently, upwelling lava filled the crater and finally spilled over the rim in broad floods. The evidence of this history is abundant. There is hardly a neck without a jacket of inwardly dipping pyroclastic debris, made up of lava and sedimentary fragments than range in size from the finest dust to blocks many yards in diameter. Normally, the dip of these ejecta increases both upward and inward. Inbedded friction breccias bordering the necks are extremely rare, and in general the enclosing sandstones and shales remain undisturbed.

Hack (1942) has found that the explosive pipes or diatremes are numerous, and that most of them occur in a dense cluster within an area of 800 square miles. He writes as follows:

In general the diameter of the vents decreases as erosion increases. In areas where dissection has been slight, at levels above the Hopi Buttes surface, explosion pits are generally 3000 to 4000 feet in diameter. In many places, the initial explosion pit is overlain by domes of lava which have pushed outward, spilling over the sides, crumpling and pushing out the underlying and bordering tuffs and sediments. In a few places these lava eruptions were of sufficient duration to form rather continuous flows. The more deeply dissected diatremes range in diameter from 500 feet to 2000 or 3000 feet. In general the material in them is less well bedded, and, if pyroclastic, is coarser.

The volcanism occurred in Pliocene time, because vertebrate bones of that age have been found in the Hopi lake beds (Bidahochi formation).

The igneous rocks of the Navajo region differ from those of the Hopi Buttes because they contain a paucity of lava flows. Also, according to Williams (1936):

. . . most of the Navajo volcanic necks are made up predominantly, not of columnar lava, but of coarse tuff-breccia and are crowded with fragments of plutonic rock, chiefly of granitic type. Petrographically, also, the two provinces differ markedly; in the Hopi Buttes, monchiquitic rocks are typical, while in the Navajo region these are far subordinate to minettes. Probably the Navajo vents were more explosive. Indeed, many of them can never have erupted lavas. How closely they resemble the well-known diatremes of the Schwabian Alb, the Rhongebirge, and Central Scotland will be apparent from what follows. Like those explosive vents, many of the Navajo volcanoes seem to be scattered at random, without regard to pre-existing structures. None is located on a fault.

Monument Valley, with its fantastic, castellated crags, is carved from the De Chelly sandstones and the Moenkopi shales that occupy the broad and gently rippled top of a domical uplift, bordered on the south and east by the sharp monocline of Comb Ridge and on the west by less-prominent folds that traverse the Rainbow Plateau. On the summit of the upwarp, dips of more than 30 degrees are rare, but in the flanking folds they may reach 60 degrees. Many intrusions are to be found along the Comb Ridge monocline, extending from the village of Kayenta in an arc to the San Juan River. These tend to follow a strong system of joints, approximately normal to the trend of the monocline. In Monument Valley itself there is much less regularity in the trend and distribution of the intrusions.

All the necks rise boldly from the surrounding sediments, despite the fact that they consist almost entirely of tuff-breccias. The few thin dikes which cut the breccias are not responsible for this resistance to erosion; it results, rather, from the compactness of the neck fillings, for the fine tuffaceous matrix has been indurated by hot solutions, and much of it is cemented by calcite. The absence of strong joints, such as cut the adjacent sandstones, is, doubtless, another contributing factor.

Shiprock, Agathla, and Alhambra Rock are some of the well-known necks referred to by Williams.

Two examples of cauldron subsidence in the Navajo region have been described (Williams, 1936), one at Buell Park, north of Fort Defiance, near the Arizona–New Mexico line, and one at Indian Wells in he Hopi Buttes. The first is about 2½ miles in diameter and collapsed at least 1000 feet. The second is ½ by ⅓ mile in diameter, and it collapsed perhaps 100 or 200 feet.

HIGH PLATEAUS OF UTAH

The High Plateaus of Utah are generally considered a subprovince of the Colorado Plateaus physiographic province. They lie along the western margin of the Colorado Plateau as shown in Fig. 26.1. The individual relief features are indicated in Fig. 26.16, and bold escarpments 2000 to 6000 feet high between valley floor and Plateau top are common.

The High Plateaus span the transition from Paleozoic miogeosyncline to shelf, and also contain the thick clastic deposits of the Cretaceous and early Tertiary described in Chapter 22 on the Central Rockies. The thick Jurassic evaporite trough also lies within them. In other words, the

Wasatch line is contained within the High Plateaus. The dotted line of Fig. 26.1 represents the boundary of the Laramide Colorado Plateau and Laramide Central Rockies as designated by Hunt (1956), and separates the comparatively flat beds of Colorado Plateau aspect from the deformed strata of the Central Rockies. However, from the southern Wasatch Mountains southward a gently deformed belt separates the highly deformed strata from the little deformed, so the boundary is not a sharp one. Also confusing the problem of boundary is the highly deformed zone in San Pete Valley between the Wasatch and Gunnison plateaus which to the writer and others seems to be a salt anticline structure. If this is considered a local structure of the Colorado Plateau and not one of regional orogenic significance, then the dividing line would lie in or west of the Gunnison Plateau.

The main structural exhibit in the High Plateaus of Utah is a normal fault zone of trenches and uplifted blocks. The faults, as well as known, are shown on Fig. 26.16. The large mid-Tertiary volcanic field occupies a central position in the High Plateaus and has been broken and offset by the faults, the same as the older sedimentary rocks. The faults are thus late Tertiary. In part they have continued active through the Quaternary because of recent earthquakes and fresh scarps in places. See Chapter 31.

The High Plateaus belong to the Basin and Range system if only the late Cenozoic normal faults are considered, but most geologists have given attention more to the Laramide structures in designating a boundary.

AGE OF UPLIFTS AND VOLCANISM

Near the close of the Cretaceous period the region probably was low (Hunt, 1956). In Late Cretaceous or early Eocene time occurred the deformation that resulted in the anticlinal uplifts such as the San Rafael Swell, Circle Cliffs uplift, and Henry Mountains structural basin. Evidence for this date of the folding is found in the St. George basin (Gardner, 1941), the vicinity of Escalante (Gregory and Moore, 1931), and the

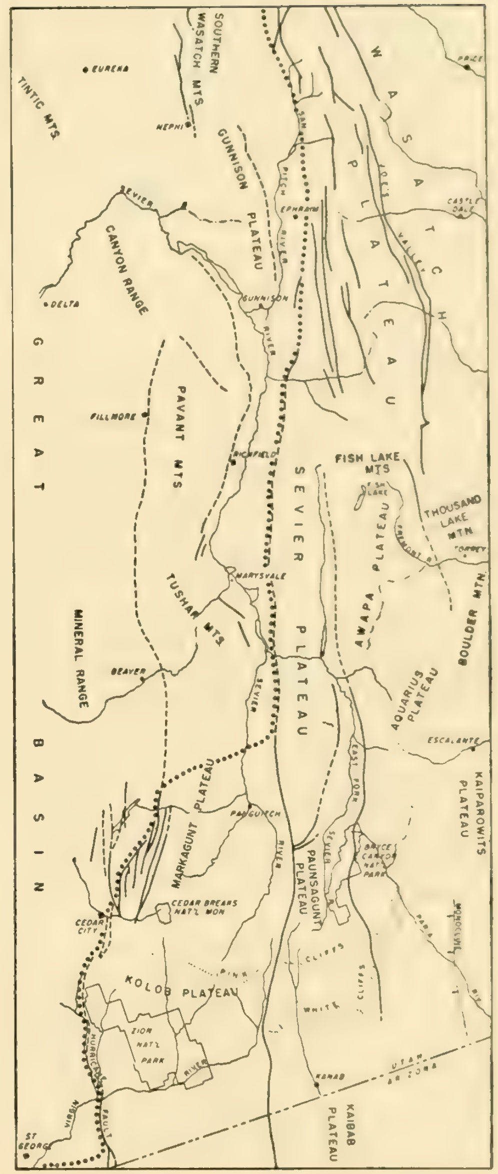


Fig. 26.16. High Plateaus of Utah. Dotted line represents western margin of Colorado Plateau according to Hunt, 1956.

north end of the Waterpocket fold (Dutton, 1880). At each of these places, Eocene strata lie across the eroded edges of the older folded rocks. It seems probable that the other large, northerly trending folds of the plateau, like the Kaibab uplift, the Defiance anticline (see Fig. 26.1), and the Monument upwarp also were formed at this time. As a first result of this folding, the structural uplifts became topographically high areas, and the structural basins became the sites of deposition of sediments eroded from the highland. San Juan basin in New Mexico and Uinta basin in Utah still preserve their basin sediments; from the other basins such fill as was deposited has been removed.

Following the monoclinal flexing and the creation of the uplifts and basins in Laramide times, the Plateau was broadly uplifted. The history of epeirogenic uplift, superposition of the major streams across the structures, and the gorge cycle of erosion are not yet well known, although considerable has been written about these fascinating subjects. Longwell (1946) reviews the problem of dating these events, and concludes provisionally that the major uplift occurred in the Pliocene. See also Hunt (1956).

The intrusion of the laccolithic clusters and the main phases of the extrusive activity in the volcanic fields probably occurred in the Pliocene. The activity can be dated accurately only in the Hopi Buttes area, but because of the similar setting of all the volcanic fields in the Colorado Plateau, it is safest to assume this age for all until proved otherwise. The volcanic activity continued in several stages down to very recent times, both north and south of the Grand Canyon and in the Datil field.

Regarding the age of the laccolithic intrusions, Hunt (1956) observes:

> The drainage on the laccolithic mountains is consequent and some of the main tributaries of the Colorado River appear to have been shifted monoclinally as a result of the doming by the intrusions. For example, the Fremont River swings in a wide arc around the north side of the Henry Mountains and follows the trough between the mountains and the San Rafael Swell. The Dolores River swings in a similar arc around the north side of La Sal Mountains and follows the trough between them and the Uncompaghre Uplift. The headward part of Glen Canyon has avoided the domes at the two southern Henry Mountains; the lower part of San Juan River and the adjoining section of the Colorado River have avoided the dome at Navajo Mountain. The adjustment of the drainage to the intrusive structures stands in striking contrast to the lack of adjustment of the drainage to the orogenic structures.
>
> This adjustment would have developed if doming of the laccolithic mountains had dammed earlier stream courses, forcing streams like the Fremont and Dolores into new courses. Inasmuch as both streams now follow the structurally lowest course possible, they may have been flowing across Tertiary basin sediments when the intrusions occurred, and their courses shifted monoclinally off the domed areas even though the doming progressed slowly.

EPEIROGENIC MOVEMENTS AND ISOSTATIC AND SEISMIC CONSIDERATIONS

The summation of movement in the Colorado Plateau in late Cretaceous, Paleocene, and Eocene time was downward around its borders, save for the south and southwest parts. The thick deposits in the High Plateaus of Utah, the Uinta and Piceance basins, and the San Juan basin attest to subsidence there, leaving the central and southern parts positive. The general relations may also be viewed as a northward down-tilting. The occurrence of gravels containing Precambrian quartzite on the Mogollon Rim (Fig. 26.9) suggests that this part of the Plateau was low-lying and was receiving debris from mountains to the south (Hunt, 1956). The several asymmetrical uplifts within the Plateau formed at this time also.

After the development of the marginal basins the entire Plateau was uplifted as a block between 6000 and 8000 feet. The uplift of the southern part was probably greater than the northern, and the impressive Mogollon Rim and Mountain Region of central Arizona was brought into existence. The epeirogenic rise is thought to be contemporaneous with the late Tertiary and Quaternary block faulting to the west with the creation of the Basin and Range province. The magmatic activity represented by the laccolithic clusters, the Hopi and Navajo volcanic fields, and the large marginal fields was also contemporaneous with the broad uplift.

For the Colorado Plateau mass of the earth's crust to stand 6000 to 8000 feet above sea level it has generally been concluded from isostatic considerations that roots of "granitic" crust several times as thick must extend downward into the subcrust. However, Tatel and Tuve (1955) believe they recognize the Moho discontinuity under the Colorado Plateau at about 30 kilometers. If so, and thus in the absence of roots, some phe-

nomenon in the mantle must be sought to explain the uplift. Expansion of the underlying column of the mantle, either due to phase changes or partial melting, or to both, seems an immediate recourse. Since magmatic activity is widespread in and around the Plateau, the writer is prone to turn to the partial melting hypothesis as the principal cause of widespread uplift. The subject is elaborated on in the final pages of Chapter 36. The asymmetrical uplifts may be great blisters incident to an early stage of megasill intrusion in the crystalline basement.

27.

SOUTHERN ARIZONA ROCKIES

PHYSIOGRAPHIC CHARACTERISTICS AND DIVISIONS

Arizona is divided into three physiographic units: the Colorado Plateau on the northeast, the Sonoran Desert region on the southwest, and the Mountain Region between them. (See index map, Fig. 27.1.) The Mountain Region and Sonoran Desert have been included in the Basin and Range physiographic province (Butler and Wilson, 1938).

The Mountain Region (also referred to as the Mexican highland) forms a belt 60 to 100 miles wide and contains most of the large ore deposits of the state. It is characterized by many short ranges nearly parallel to each other and to the margin of the Plateau. The individual ranges are separated by valleys deeply filled with fluviatile and lacustrine deposits which are now generally being eroded to widespread pediments capped by veneers of gravel.

The southern margin of the Colorado Plateau is usually taken as the erosional escarpment of the Kaibab limestone and Coconina sandstone, called the Mogollon Rim, but lower Paleozoic beds extend southward in certain summit areas well within the Mountain Region. The Mogollon Rim is at an altitude of 8000 feet or slightly less, and Phoenix near the junction of the Mountain Region and the Sonoran Desert is 1100 feet above sea level. The elevation of Tucson is 2372 feet and Bisbee in the Mountain belt is 5490 feet. The Sonoran Desert (also called the desert region) is characterized by a great preponderance of broad desert plains over mountain ranges. The ranges are relatively short and far apart, and generally have lower elevations than those of the Mountain Region.

According to Ransome (1933):

> Probably the most impressive feature of the landscape, to one who sees it for the first time, is the sharp contrast between steep and rugged mountains and wide expanses of desert plain. It is true that the plains merge imperceptibly with smooth, evenly graded alluvial slopes, which may attain considerable altitude where they meet the mountain fronts, but the presence of these great ramps of detritus scarcely detracts from the general striking contrast between mountain and lowland. Such topography is, of course, characteristic of most mountainous desert regions, in which the greater part of the debris washed from the mountains is deposited in the adjacent valleys, these gaining in extent and becoming more plainlike as the minor eminences are worrn down and buried.

The surface forms of the Sonoran Desert may be classified into three groups, according to Gilluly (1937c), which are:

> . . . (1) The mountains, commonly rugged and steep-sided, with either bare rock at the surface or only a thin cover of talus; (2) the pediments, smooth carved-rock plains that generally border the mountains and are strewn with a thin but discontinuous mantle of gravel; and (3) the bajadas, smoothly rounded alluvial aprons that slope forward into the axes of the "valleys." Of these, the mountains and pediments are chiefly carved by erosion; the bajadas are chiefly depositional.

A glance at the *Geologic Map of the United States* will show that the ranges of the southern California and Arizona and southwestern New Mexico are smaller, more irregular in shape, less linear and parallel, and

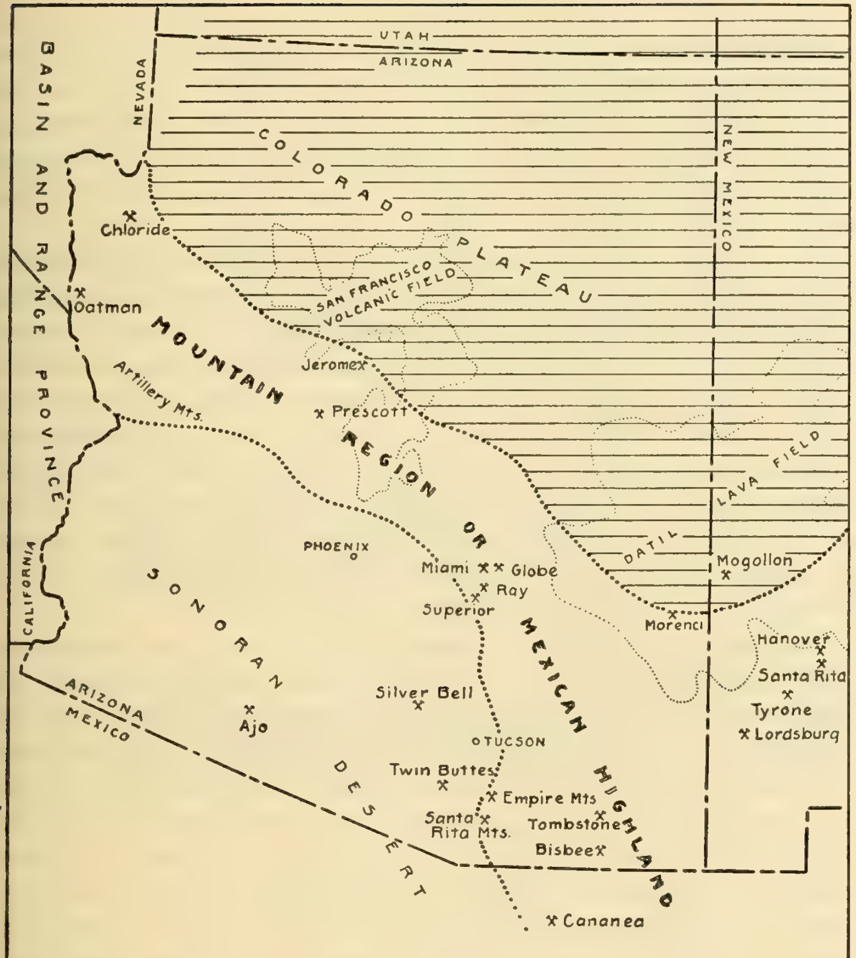


Fig. 27.1. Index map of Arizona showing the central mountain region and the desert region (Sonoran Desert). Both mountains and desert regions are part of the Basin and Range physiographic province. The chief mining districts are also shown together with mountains for which cross sections are given in Figs. 27.6, 27.7, and 27.9.

separated by relatively wider basins than those of western Utah and Nevada. Hence, the inclusion of the Sonoran Desert of Arizona in the Basin and Range province from a structural point of view must be made with reservations. The crisp boundaries imparted to ranges by block faulting are generally absent, and if the region is one of extensive block faulting, then the faults are older than those in Utah and Nevada, and erosion has beaten the fault scarps back considerable distances to form broad flanking pediments. The desert floors are in need of gravity surveys to delimit buried faults and their patterns, if such exist. Some reports refer to late Tertiary normal faults as Basin and Range faults, but other reports do not use the term Basin and Range.

PALEOZOIC AND MESOZOIC BASINS

The paleotectonic maps of Figs. 5.1 through 5.8 show that the Transcontinental Arch dominated most of Arizona in Paleozoic time. Except in the southeastern corner of the state (Figs. 27.2 and 27.3), the deposits are thin or absent, and it does not seem possible that they could have affected in any major way the pattern of later Mesozoic and Tertiary structures.

Triassic and Jurassic sedimentary rocks are absent in the Mountain Region and Sonoran Desert and hence events which may have occurred during these times cannot be accurately dated stratigraphically. Lower Cretaceous rocks are present in southeastern Arizona as part of the Mexican geosyncline, and most igneous and deformational events there can be dated either as pre-Lower Cretaceous or post-Lower Cretaceous. Late Cretaceous strata also occur in places there, and assist further in dating of events. However, over most of the Sonoran Desert and Mountain Region, outside of this southeastern part of the state, the ages of rock masses and structures are poorly known. The history is eventful, and the sequence of events can be established but few of them accurately dated.

The Gila conglomerate whose oldest fossils to date are early Pliocene (Anderson *et al.*, 1958), but which for the most part probably is late Pliocene (Knechtel, 1936), is widespread, and serves as an upper dating plane. Events between the Cretaceous (generally Lower Cretaceous) and the Pliocene must be spaced or interpolated according to the best judgment of the researchers concerned.

USE OF TERMS, LARAMIDE AND NEVADAN OROGENIES

Because of the inability of geologists to date accurately events in southern Arizona during the Mesozoic and Cenozoic, the terms Nevadan

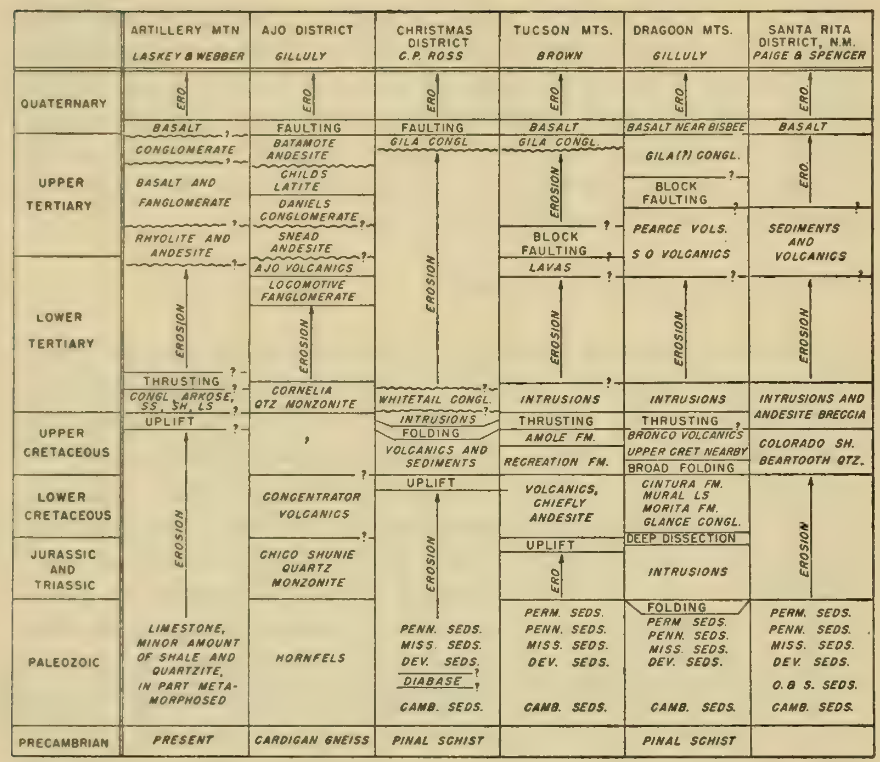


Fig. 27.2. Comparative histories of districts in the Mountain Region and Sonoran Desert of Arizona. The question marks indicate uncertainty of age assignments, but the sequence of events is fairly secure.

orogeny, Laramide orogeny, and Basin and Range orogeny are not used by authors of some of the most authoritative works. Where used in this chapter the attempt is made to make clear the uncertainties involved.

MESOZOIC AND CENOZOIC GEOLOGY OF SOUTHEASTERN ARIZONA

The ranges of southeastern Arizona (Fig. 27.4) contain Paleozoic strata representative of the Cambrian, Devonian, Mississippian, Pennsylvanian, and Permian periods, and also a thick succession of Lower and Upper Cretaceous formations. These either do not occur or occur in limited or altered form in other parts of the Mountain and Desert regions, and hence, the nature of crustal deformation and igneous activity is better recorded and deciphered in southeastern Arizona than in the south-central or southwestern part of the state. It is best, therefore, to refer to this region first for an understanding of the Mesozoic and Cenozoic geology before turning to the other areas.

The Paleozoic section including Permian beds described for the Bisbee district in Fig. 27.3, is characteristic of southeastern Arizona and adjacent New Mexico. These beds were fairly sharply folded some time after the Permian and before the deposition of the overlying Lower Cretaceous beds (Bisbee group). Examples of the folds, thrusts, and unconformable relations are shown in Fig. 27.5 and 27.6. After the folding and faulting and before the deposition of the basal Glance conglomerate of the Bisbee group a number of plutons were intruded including the Gleeson quartz monzonite, the Copper Belle monzonite porphyry, the Turquoise granite, the Juniper Flat granite, and the Cochise Peak quartz monzonite of the Dragoon and Mule mountains (Gilluly, 1956).

The Mexican geosyncline extended northwestward into southeastern Arizona and southwestern New Mexico, and in the Bisbee district the basal beds of Comanche age are conglomerates (Glance conglomerate) which range up to 500 feet thick. Overlying sandstones, shales, and limestones attain great thickness, estimates of which range from 5000 to 18,000 feet. Fossils collected from the Mural limestone, about 2500 feet above the base of the Comanche series, are Trinity in age. At a number of other localities in southeastern Arizona, such as Tombstone and in the Huachuca, Patagonia, Oro Blanco, Baboquivari, Sierrita, Tucson, Santa Rita, Empire, and Whetstone Mountains, masses of sediments, presumable Comanchean, rest unconformably upon Paleozoic or older rocks (Ransome, 1933). See the cross sections of Fig. 27.6. In places a deeply dissected surface was buried by the Lower Cretaceous sediments. The intrusive rocks had been exposed in this erosion cycle (Gilluly, 1956).

If the correlations of the table, Fig. 27.2 are correct, then the Lower Cretaceous Mexican geosyncline was continued to the northwest chiefly by a volcanic fill. The volcanic series of the Tucson Mountains (Brown,

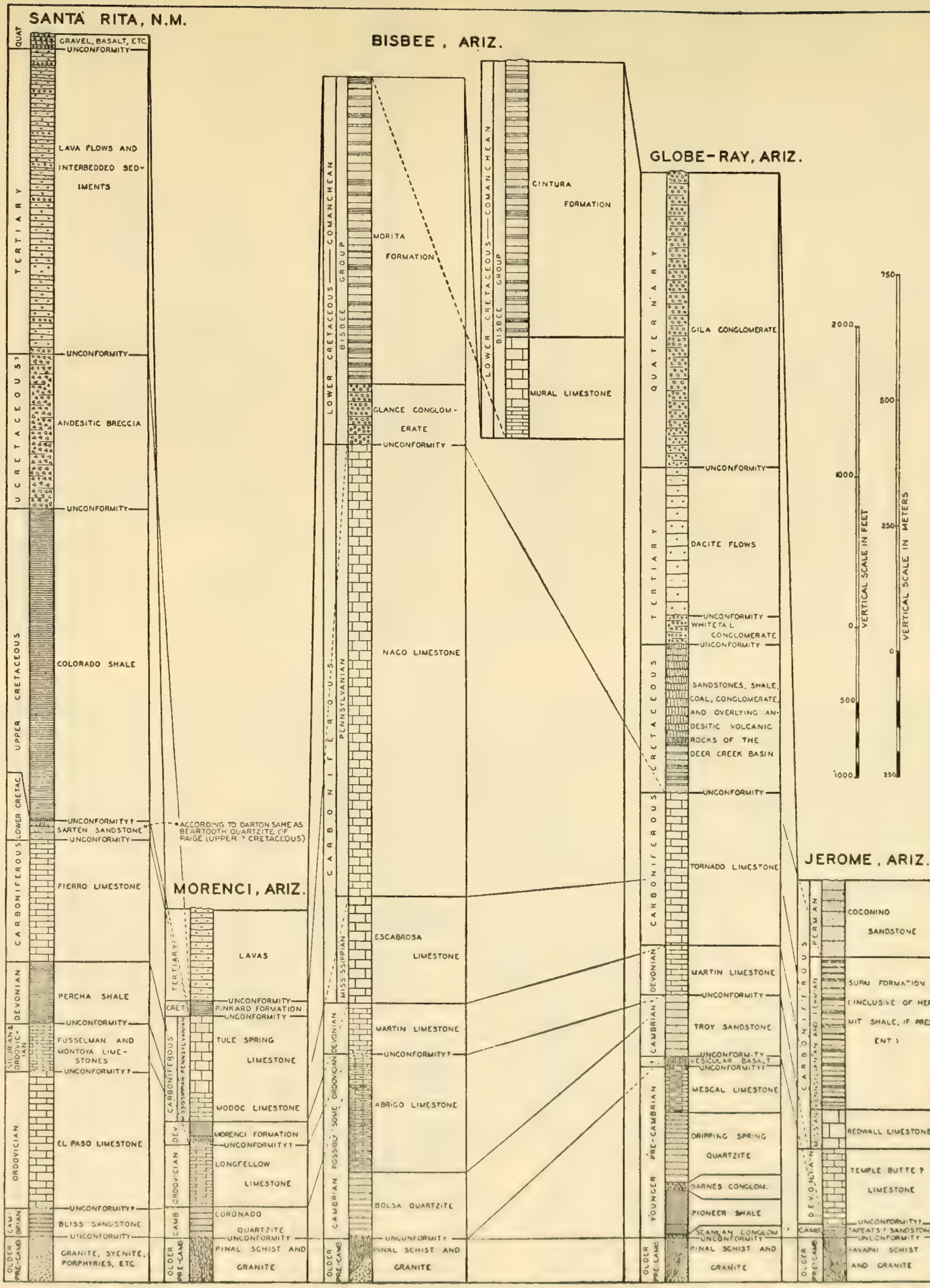


Fig. 27.3. Stratigraphic sections in central and southeastern Arizona. After Ransome, *Guidebook* 14, 1933.

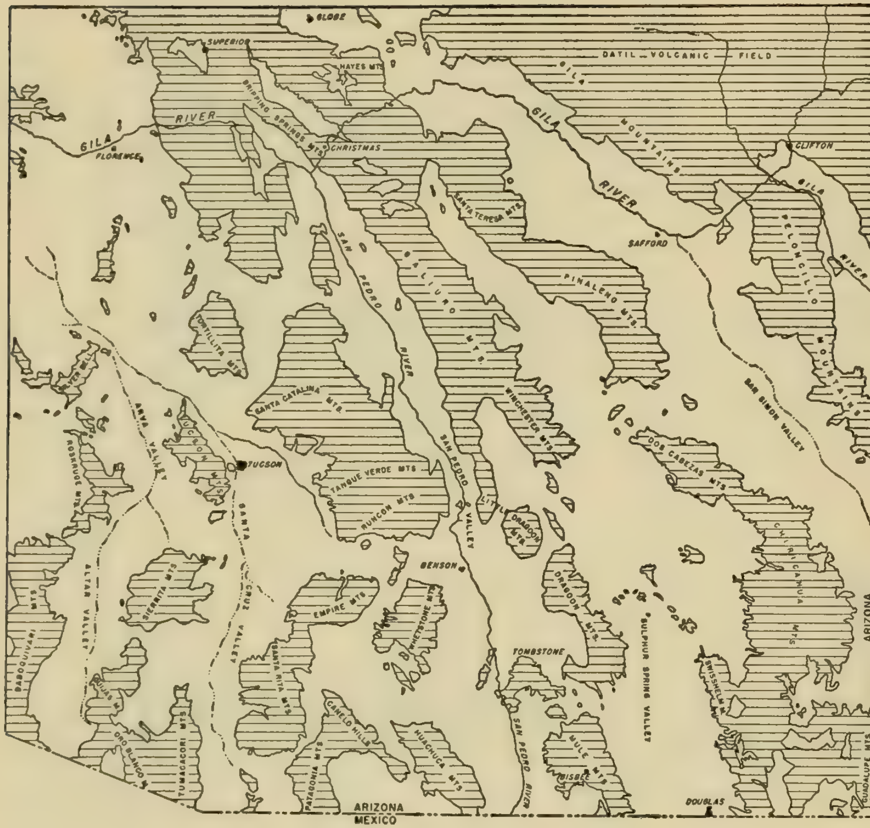


Fig. 27.4. Mountains of southeastern Arizona.

1939) composed of rhyolite, andesite, tuff, and arkose is considered Lower Cretaceous here; some of the andesites of the Christmas area (Ross, 1925) may be Lower Cretaceous, and the Concentrator volcanics of the Ajo district (Gilluly, 1946) are of the same age evidently. These are chiefly andesites and keratophyres. It was a belt of andesitic eruptions, chiefly, both of the explosive and passive kinds of activity.

Upper Cretaceous strata have not been recognized in as many places in southeastern Arizona and southwestern New Mexico as the Lower Cretaceous, nor are they as thick where known; but Reeside's map (1944) shows them to have once covered the entire area. At all places observed, there is a marked unconformity at the base (Ransome, 1933).

Because the Upper Cretaceous beds, where identified, rest in angular unconformity on all the older rocks, it is concluded that crustal movements of some proportions occurred about at the close of the Early Cretaceous. Reeside (1944) shows central and southern Arizona out of water during all but the last of Late Cretaceous time (Fox Hills and Lance time), when a trough in the site of the earlier one formed and sank at least 7500 feet. It seems possible that the sediments could come from the geanticlinal area on the southwest, but the spread of data does not preclude the existence of land areas to the northeast of the trough. Ross (1925) believes a shore line was immediately west of the Christmas–Ray–Miami districts, and this is shown on the Late Cretaceous paleotectonic map as a narrow volcanic peninsula extending southeastward from the main geanticlinal area. A northwest trend to the structures of central and southern Arizona had thus become established.

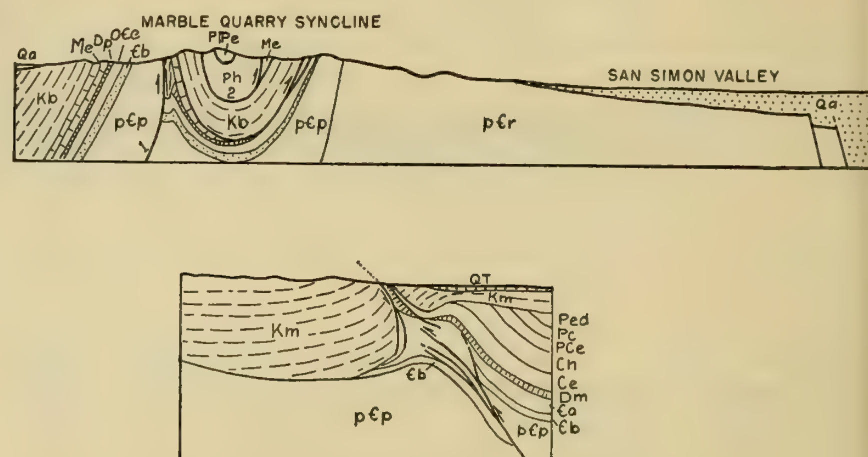


Fig. 27.5. Section through Chiricahua Mountains showing Apache Pass fault (1) and Fort Bowie thrust (2), after Sabins, 1957. Compare with Fig. 27.8.

Section in the Mule Mountains showing angular unconformity between Lower Cretaceous Morita fm (Km) and Paleozoic formations; also post-Lower Cretaceous thrusting. After Gilluly, 1956.

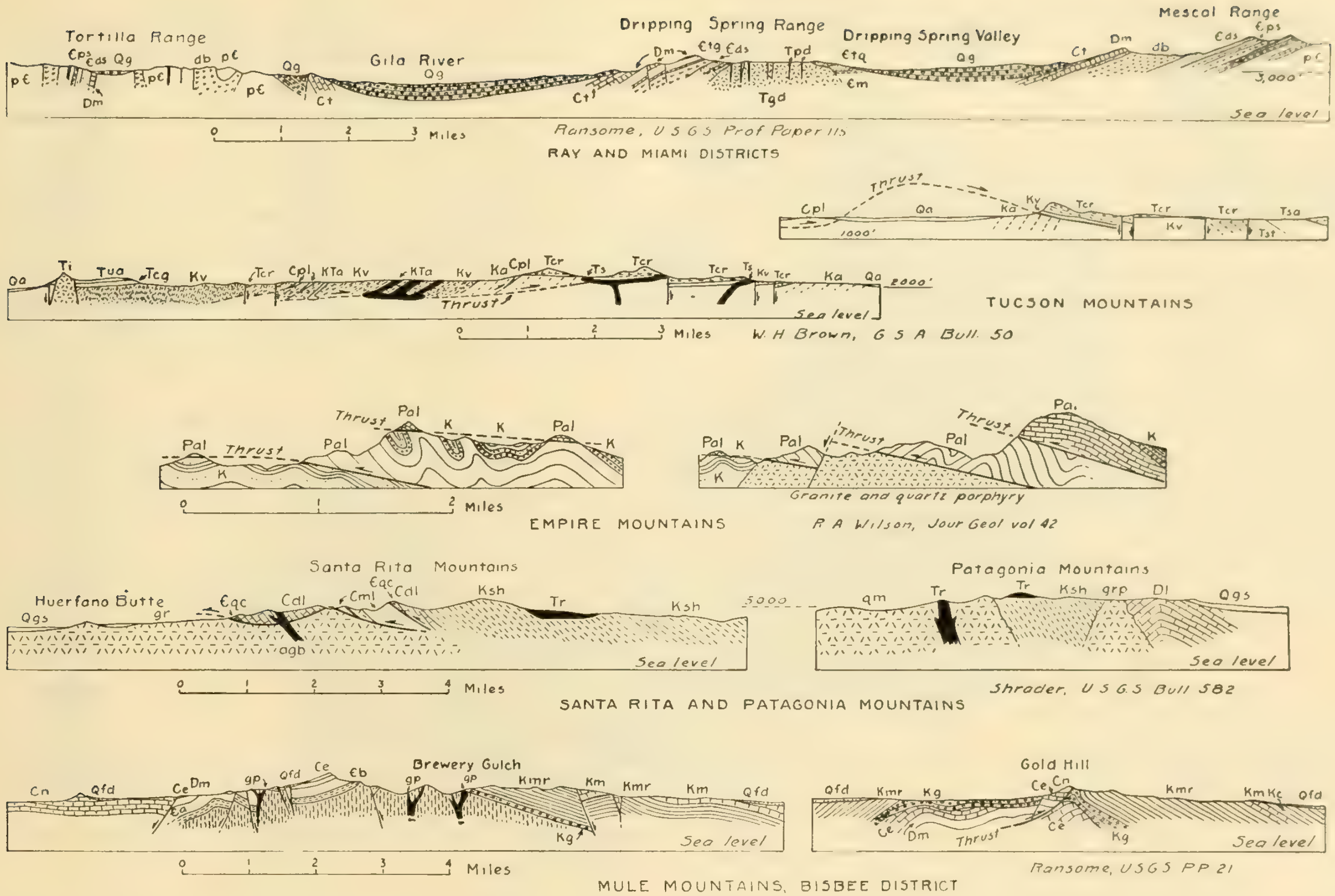


Fig. 27.6. Cross sections in southeastern Arizona. Symbols for Ray and Miami districts; Єps, Pioneer sh.; Єds, Dripping Spring quartzite; Єm, Mescal ls.; Єtq, Troy quartzite; Dm, Martin ls.; Ct, Tornado ls; Qg, Gila conglomerate, db, Mesozoic (?) diabase; Tgd, granodiorite; Tpd, quartz diorite. Tucson Mountains; Cpl, Permian ls; Kv, Cretaceous volcanic rocks; Ka, Amole arkose; Ti, acid dikes and volcanic necks; Tua, upper andesite; Teg, conglomerate; Tcr, Cat Mt. rhyolite; KTa, Amole latite; Ts, spherulitic rhyolite; Qa, alluvium and talus. Santa Rita Mountains, Єpc, quartzite, and congl.; Dl, Devonian ls.; Cml, light ls.; Cdl, dark ls.; Ksh, chiefly Mesozoic but may include some Precambrian; qm, quartz monzonite; grp, granite porphyry, agh, alaskite granite porphyry; Tr, rhyolite. Mule Mountains; ps, Pinal schist; Єb, Bolsa quartzite; Ca, Albrigo ls.; Dm, Martin ls; Ce, Escabrosa ls.; Cn, Naco ls.; Kg, Glance congl.; Kmr, Morita fm.; Km, Mural ls.; Kc, Cintura fm.; Qfd, fluviatile deposits.

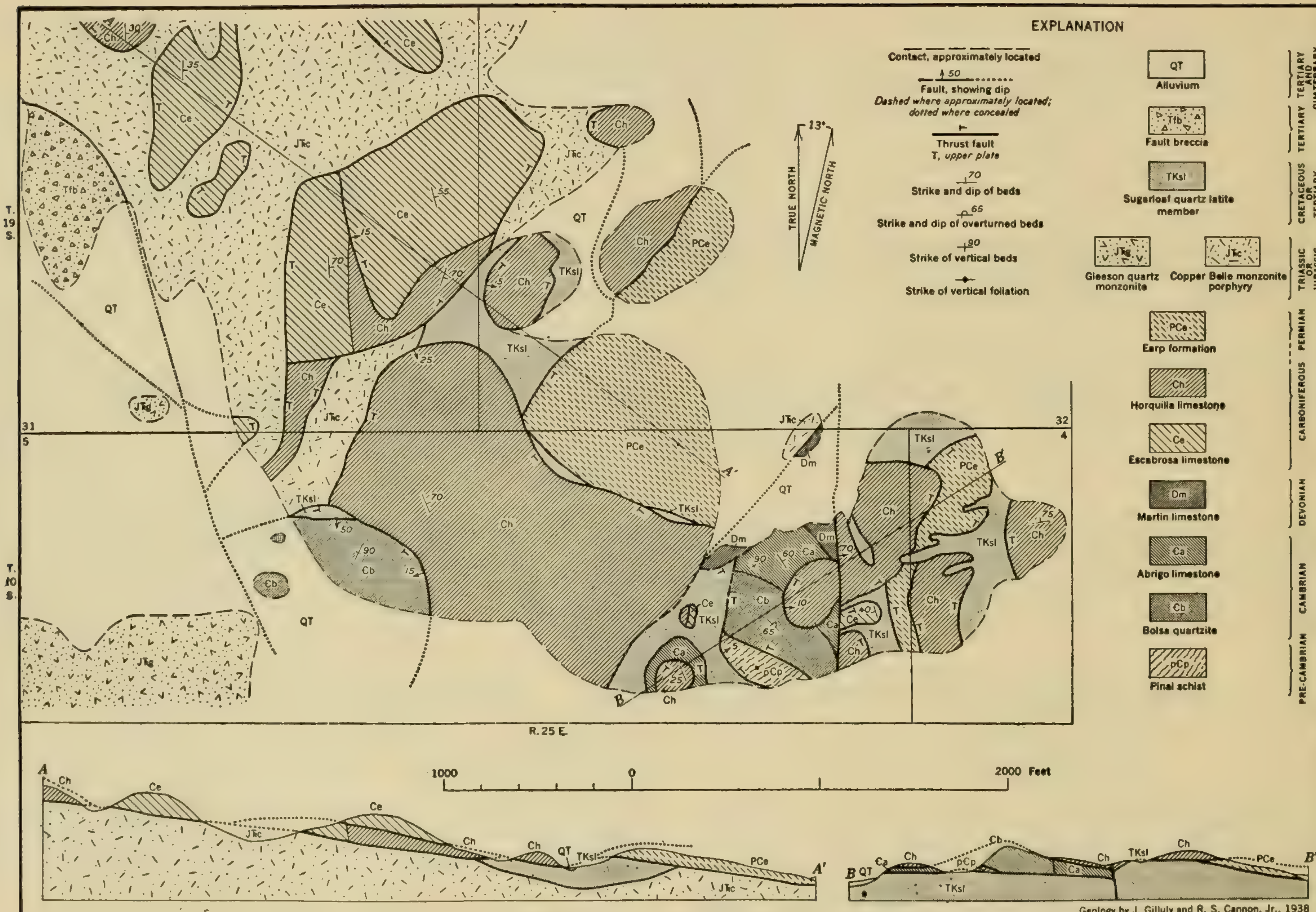


Fig. 27.7. Relationships in the thrust breccia east of Gleeson, Dragoon Mountains. Reproduced from Gilluly, 1956.

In the northern Dragoon Mountains andesite and quartz latite were erupted on an erosion surface of mild relief developed after the deposition of the Bisbee sediments. These volcanics may be Late Cretaceous in age; at least they preceded the strong thrusting.

> The most profound deformation of the area took place after the Bronco volcanics and Sugarloaf quartz latite were erupted. This involved great thrust faults of northerly to northwesterly trend in the Dragoon Mountains and the overturning of a section of the Bisbee formation fully 3 miles thick along the eastern flank of this range. A gigantic breccia of fragments of nearly every older formation exists in the Courtland and Gleeson areas. Refer to Fig. 26.6. It suggests that the major fault was, in this section of its exposed course, advancing over the surface, producing the breccia by attrition of the overriding thrust plate. Minor thrust fragments of this age are found in the Tombstone Hills. . . .
>
> The Stronghold granite is younger than the thrusting and has domed the thrust sheets slightly. This doming does not appear, however, to account for the emplacement of the granite, which is clearly transgressive. In the Tombstone Hills the Schieffelin granodiorite seems also to be younger than all important compressional stresses, as is the Uncle Sam porphyry.

The sequence and pattern of Laramide (?) events in the Chiricahua Mountains is instructive. The northwest course of the main structures dominates the geologic map, whereas the giant thrust breccia in the Dragoon Mountains leaves the trends uncertainly recognized there. According to Sabins (refer to Fig. 27.8),

> During the major post-Comanche to pre-Pliocene orogeny, strong southerly to southwesterly horizontal compression caused the following tectonic sequence. The autochthonous rocks along the northeast front of the range were overridden from the southwest by the first thrust sheet. Strike-slip displacement along the Emigrant fault cut the autochthonous block and the overlying thrust sheet, which was separated into the Fort Bowie plate and the Wood Mountain plate. The Fort Bowie plate was later folded to form the Marble Quarry syncline and was truncated by the younger Fort Apache reverse fault. Finally, the Whitetail plate overrode the Fort Apache fault block.

Volcanic extrusions of approximately mid-Tertiary age then accumulated on an erosion surface on the deformed strata. These were faulted and tilted, possibly just prior to the deposition of the Gila conglomerate in Pliocene time. Not only was the faulting the cause of the deposition of the Gila conglomerate, but also probably the modern ranges were blocked out by it at this time. Some faulting continued after the Gila conglomerate accumulated.

The age of the thrusting cannot be more accurately placed than in the Late Cretaceous (probably very Late Cretaceous) or Early to Mid-Tertiary, but when it and the stocks of post-thrusting age are compared to a similar sequence of events in Utah, Nevada, Colorado, and central New Mexico, one may logically point to a Laramide age. The folding of the Upper Cretaceous beds of the Mexican geosyncline in Coahuila, Chihuahua, and Sonora is generally referred to as Laramide, and this broad region continues the folding and thrusting of southeastern Arizona and southwestern New Mexico southward.

The picture described in the above paragraphs of the structure of the ranges of southeastern Arizona is clarified by Jones (1961), who recognizes most of the ranges to be complex anticlines with Precambrian or Triassic–Jurassic granite in their cores. The structural relief of some of the uplifts is 25,000 feet. Some began to rise in Mesozoic time and continued intermittently through at least the Miocene. High angle reverse faults define the flanks, and appreciably downslope mass movement has occurred to form the low-angle thrusts.

MESOZOIC AND CENOZOIC GEOLOGY OF SOUTHERN ARIZONA

The geology of the Ajo mining district of south-central Arizona in the Sonoran desert is well known from the work of Gilluly (1946), and probably is representative of the geology of this general region. The main rock units and events are listed in Fig. 27.2. A cross section is shown in Fig. 27.9. The only fossils found in the entire district are in blocks of limestone in the Locomotive fanglomerate, presumably of about middle Tertiary age. The fossiliferous boulders are referable to the Devonian, Mississippian, and Pennsylvanian, and hence when the fanglomerate was being deposited outcrops of beds of these Paleozoic ages were probably nearby in upland areas. The rocks and events, although their sequence is relatively well established, are not dated by stratigraphic methods, and the ages assigned are very tentative.

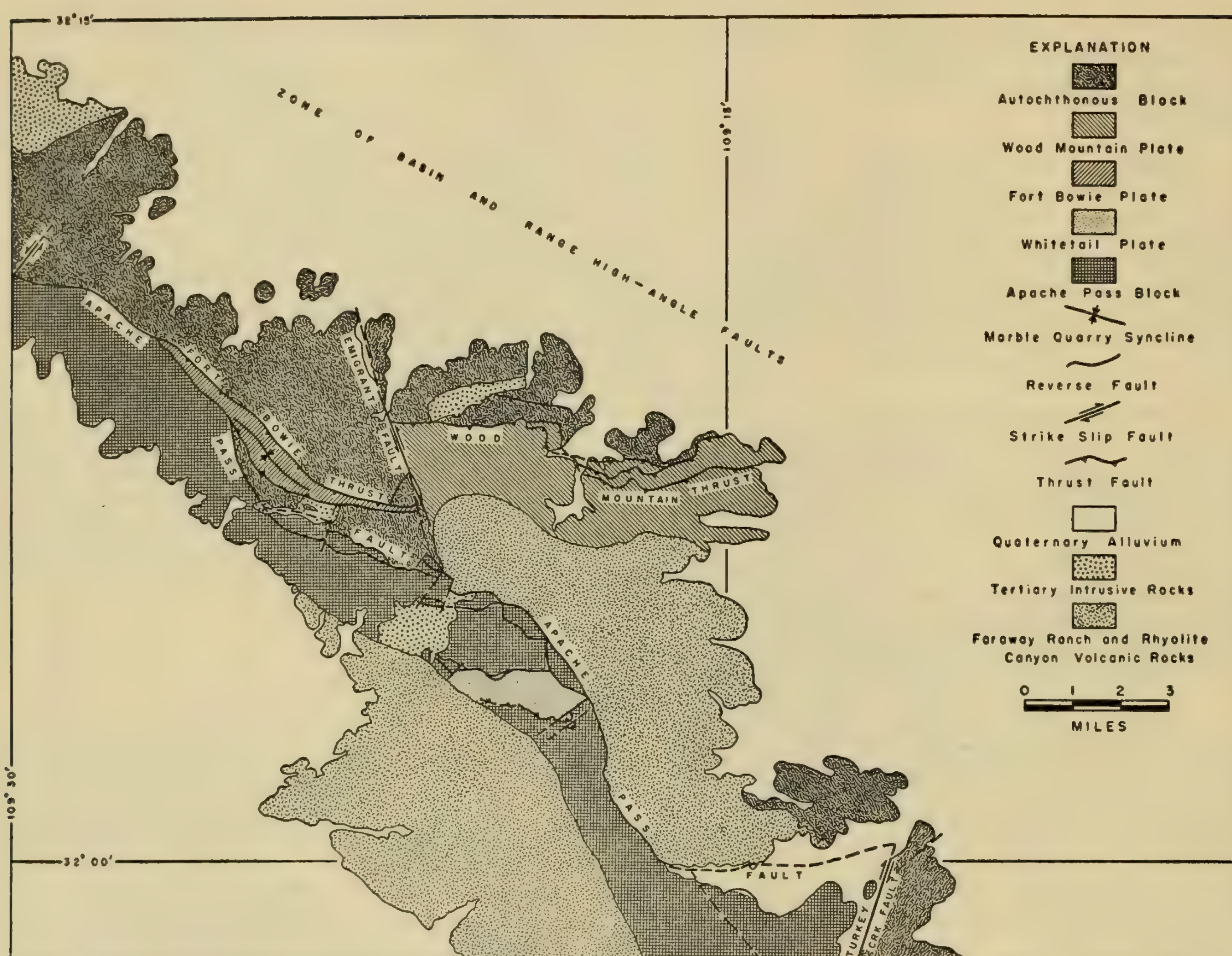


Fig. 27.8. Thrust plates of the Cochise Head and Vanar quadrangles, Chiricahua Mountains. Reproduced from Sabins, 1957. Compare with Fig. 27.5.

Of particular importance in regional tectonics are the ages of the Chico Shunie quartz monzonite and the Cornelia quartz monzonite. The Chico Shunie precedes the Concentrator volcanics and the Cornelia follows. The sequence is similar to the one in the Dragoon and Mule Mountains of southeastern Arizona, except for the absence of the Bisbee sediments, which may be represented in part by the Concentrator volcanics. Gilluly assigns the Chico Shunie intrusion questionably to the Mesozoic and the Cornelia intrusion to the early Teritary. Speculatively,

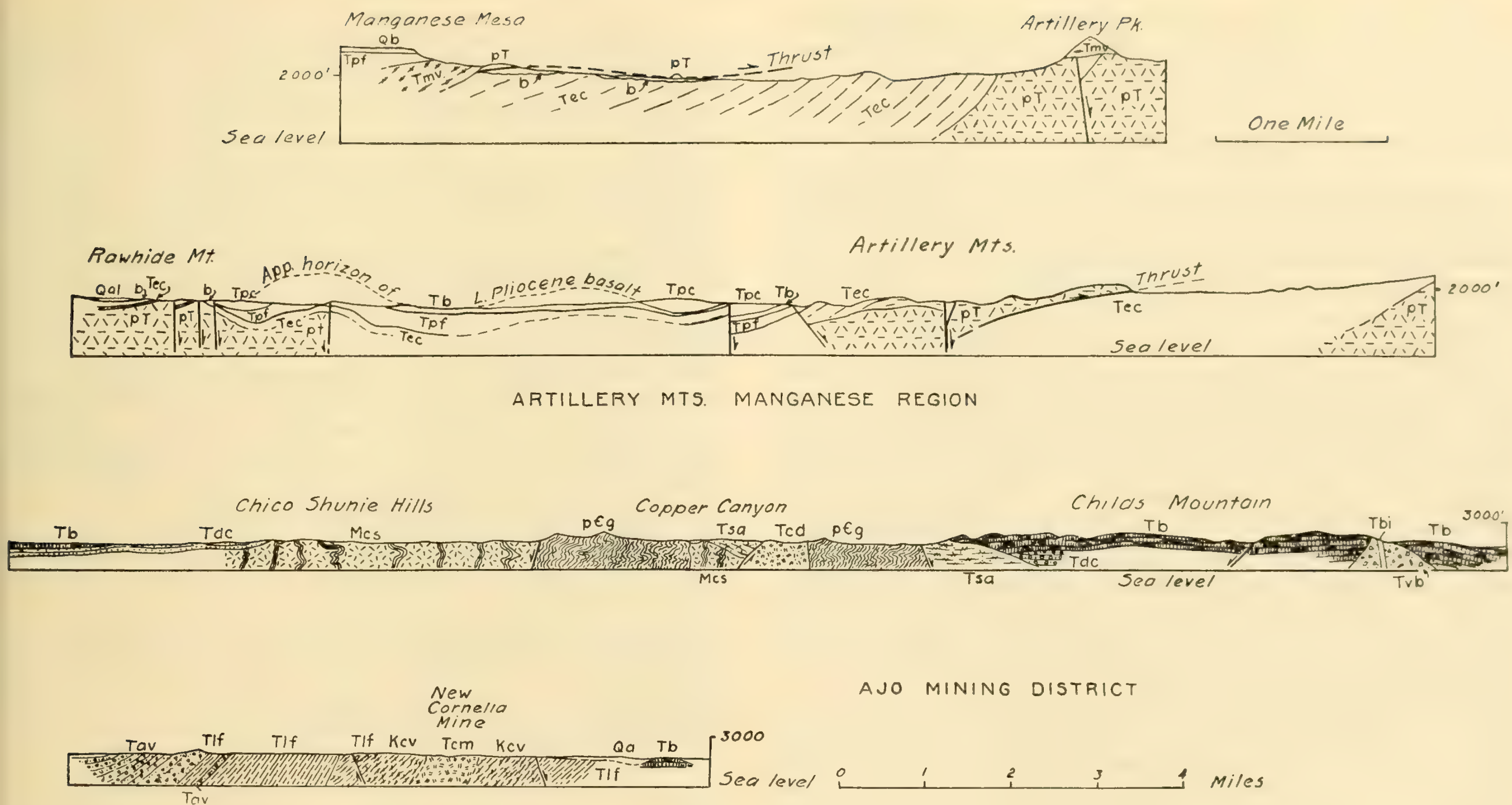


Fig. 27.9. Cross sections of the Artillery Mountains, after Lasky and Webber, 1944, and the Little Ajo Mountains, after Gilluly, 1937a.

Formations in the Artillery Mts.; pT, Paleozoic ls., sh., and qutz., and Precambrian granite, gneiss, and schist; Tec, lower Eocene (?) congl., ss., and sh.; Tmv, Miocene (?) volcanic rocks; Tpf, lower Pliocene (?) alluvial fan and playa deposits containing manganiferous beds; Tb, lower Pliocene (?) basalt; Tpc, upper Pliocene (?) basalt; Qb, Early Pleistocene (?) basalt.

Formations in the Ajo mining district; p€g, Cardigan gneiss; Msc, Chico Shunie quartz monzonite; Kcv, Concentrator volcanics; Ted, Cornelia quartz monzonite, Tlf, early Tertiary Locomotive fanglomerate; Tav, middle (?) Tertiary Ajo volcanics; Tsa, middle (?) Tertiary Sneed andesite; Tdc, middle (?) Tertiary Daniels conglomerate; Tcl, Pliocene (?) Childs latite; Tvb, Pliocene andesite breccia; Tb, Pliocene andesite flows; Tbi, intrusive basalt.

the Chico Shunie could correlate in age with the Juniper Flat granite and other related plutons, and the Cornelia with the Stronghold granite and Schieffelin granodiorite.

GEOLOGY OF WEST-CENTRAL ARIZONA

The Artillery Mountains near the west end of the mountain region has been studied by Lasky and Webber (1944), and the rock units and succession of events there are probably representative of the general area. The stratigraphy is briefly described below, and the relation of rock units to the structural events is shown in Figs. 27.2 and 27.9.

Of particular interest are two thrust sheets, one of which occurs in the Artillery Mountains and one in the Chemehuevis Mountains just southeast of Needles (see *Tectonic Map of the United States*). Both sheets overrode toward the Colorado Plateau and are probably Laramide in age.

The Laramide orogeny seems to have been in two phases, first an uplift that furnished the lower Eocene (?) conglomerate, arkose, sandstone, shale, and limestone beds to a northwestward-trending trough, and then the thrusting that brought the Precambrian and Paleozoic (?) sedimentary rocks on top of the lower Eocene (?) beds. These lower Eocene beds have a structural setting similar to the conglomerates, arkosic sandstones, and claystones of the New Water Mountains farther south.

The Artillery Mountains thrust is overlapped by the Miocene (?) volcanics. Minor normal faulting then formed a graben in which lower Pliocene (?) sediments and an overlying basalt accumulated. After the graben basin had become integrated into a regional drainage system, the upper Pliocene (?) conglomerate was deposited. The Pliocene (?) rocks were then folded into a shallow composite syncline that parallels the northwestward trend of the basin, and that now occupies the valley between the Artillery and Rawhide Mountains.

The folded rocks along either side of the valley, together with the overlying Pleistocene (?) basalt, are broken by northwestward-trending normal faults, which presumably are the effect of renewed movement along older fault zones. See cross sections of Fig. 27.9.

Rock Units of the Artillery Mountains Manganese Area	Thickness, Feet
Recent:	
Talus deposits, and gravel and sand along the present drainage.	
Erosional Unconformity	
Later Pleistocene:	
Pediment gravel and valley fill.	
Angular Unconformity	
Earlier Pleistocene (?):	
Massive, fine-grained to vesicular glassy basalt.	0–350 plus
Angular Unconformity	
Upper Pliocene (?):	
Largely light to dark red, poorly sorted conglomerate with discontinuous bedding. Includes a prominent basalt member in the southwestern part of the area.	0–2000 plus or minus
Erosional Unconformity	
Lower Pliocene (?):	
Massive aphanitic vesicular basalt	0–250 plus
Alluvial fan and playa deposits—fan-glomerate, conglomerate, sandstone, siltstone, mudstone, clay, and limestone; in part gypsiferous. The principal manganese-bearing formation.	0–1500 plus or minus
Angular Unconformity	
Miocene (?):	
Tuffs, breccias, and flows, rhyolitic to andesitic.	1800 plus
Angular Unconformity	
Lower Eocene (?):	
Conglomerate, arkose, sandstone, shale, limestone, and a little clay, with some tuff and a widespread basalt member; in large part highly indurated.	2500 plus
Angular Unconformity	
Paleozoic (?):	
Limestone with minor quantities of shale and quartzite in part metamorphosed.	
Angular Unconformity	
Precambrian:	
Granite, gneiss, microbreccia, and subordinate schist, including some monzonitic rock in the Rawhide and Buckskin Mountains that may be of post-Cambrian age.	

NEVADAN OROGENY (?)

Post-Permian and pre-Lower Cretaceous folding is recorded in southeastern Arizona and to the southeast in Coahuila (see Chapter 14). In both places granitic magmas have intruded the folded strata and were exposed by erosion before the Lower Cretaceous beds were deposited. To the west at Ajo, folding (?) and metamorphism preceded the intrusion of the Chico Shunie quartz monzonite. These events seem to correlate with the pre-Lower Cretaceous orogeny in southeastern Arizona. Farther to the northwest in the Artillery Mountains Paleozoic limestone, shale, and sandstone were in part metamorphosed before the Tertiary, at least. No intrusions of possible Mesozoic age are noted there, however. Then in central and western Nevada a long succession of deformational events are documented from late Devonian to the close of Jurassic time. In the Kimmeridgian (latest Jurassic) considerable volumes of granitic magma invaded the folded and thrust-faulted strata. This Late Jurassic orogeny has been classed as early Nevadan in Chapter 17.

It will be recalled (Chapter 14) that the Coahuila peninsula rose in Kimmeridgian time, and that the Mexican geosyncline took form to the west during the Late Jurassic. It received sediments from a geanticlinal area on the west as well as from the Coahuila peninsula. The rise of the peninsula and geanticline may indicate that both were orogenic belts. This was a time of thrusting and intrusions in central and western Nevada. These coincident relations support the thesis that southern Arizona was a belt of orogeny in the Late Jurassic and that the belt existed as a branch from the main belt in Nevada which continued southward into Sonora, Mexico, in the region of the geanticline that lay west of the Mexican geosyncline. See tectonic map of Plate 10.

IGNEOUS CYCLES AND MINERALIZATION

Igneous rocks possibly of Palezoic age have been described in two places. According to Ettlinger (1928), a diabase in the mountain region of central Arizona is intrusive as multiple sills in Cambrian strata but not in any younger Paleozoic strata and, therefore, may be pre-Devonian. The diabase extends over 1600 square miles, and the combined thickness of the sills in places approaches a mile. Others have suggested a post-Permian and Cretaceous age. It predates the mineralization of the region. Gilluly (1946) regards a hornfels in the Ajo district of the desert region as possibly Paleozoic. It, however, is of andesitic and rhyolitic derivation, unlike the composition of diabase.

Butler and Wilson (1938) list the Juniper Flat stock of the Bisbee district as post-Paleozoic and pre-Cretaceous, and suggest that the activity may be Nevadan in age. Ransome considered the Sacramento stock at Bisbee and associated metallization also as pre-Cretaceous. As already noted, Gilluly identifies several plutons in the Dragoon Mountains as post-Permian pre-Lower Cretaceous.

A large number of stocks that range from granite to diorite occur in the mountain and desert regions of Arizona, and all are probably younger than the Kaibab (Permian) limestone. Of late years they have been considered Laramide in age, mainly upon the argument that they are similar in lithology and their structural setting is similar to other known Laramide intrusions of the southwest. Definite proof of Late Cretaceous age is probably not obtainable for most of them.

A group of the stocks in central Arizona are considered to be cupolas of a major underlying parent pluton called the central Arizona batholith (Ettlinger, 1928).

In general, the stocks have little plan or pattern in their distribution. In the Superior–Miami–Globe and Morenci–Metcalf districts, however, the intrusive bodies have a general northeastward direction across the mountain region. Likewise, the ore-bearing fissures in these and other districts strike northeastward (Butler, 1938). The mining districts shown on the index map, Fig. 27.2, are a pretty good clue of the distribution of the stocks. Many of the ore deposits of Arizona are due to the mineralizing activity of the magmas of these stocks, especially in the central and southeastern part of the state.

Lava outpourings are very extensive in Arizona and New Mexico, and some are shown on the index map of Fig. 27.2. Refer also to the *Geologic* and *Tectonic* maps of the United States. In part they are younger than the monzonitic intrusions, but in part they are possibly contemporaneous or

even older. The most extensive and probably thickest is the Datil field, which is described in Chapter 36. In the mountain region of Arizona, the Laramide (?) stocks had been exposed by erosion and then were covered unconformably by the lavas. The Tertiary in general was a time of prolonged volcanic activity from place to place, the lavas were more acid than those of the Cretaceous, and widespread block faulting was prevalent. See charts, Figs. 27.2 and 27.3.

A third group of mineral deposits is associated with the Tertiary lavas. The districts that belong to this class are listed in the following table. The ore deposits are in the form of fault veins that cut the lavas. The veins are generally crustified, shallow in depth, and contain adularia. Gold is the chief ore mineral.

Age of Ore Deposits in Arizona and New Mexico

Precambrian	Nevadan	Laramide	Late Tertiary
Jerome-Prescott	Bisbee?	Magma	Mogollon
Pecos	Patagonia?	Globe	Steeple Rock
	Red Bed copper deposits?	Miami	Lordsbury?
		Ray	Stanley Butte
		Christmas	Oatman?
		Morenci	Ajo?
		Tombstone?	Silver Bell?
		Twin Buttes	
		Magdalena	
		Santa Rita-Fierro	
		Pinos Altos	
		Tyrone	
		85 Mine?	
		Silver-manganese metallization below shale beds in southwest New Mexico; Silver City, Cooks Peak, Kingston, etc.	

TERTIARY NORMAL FAULTING

Everywhere, it seems, in the mountain and desert regions of Arizona, high-angle faults cut and offset the bedrock. They trend in many directions. They both predate and postdate the Gila conglomerate of Pliocene–Pleistocene age; some predate the Laramide orogeny, some are part of it, but the majority postdate it, and are Middle and Late Tertiary.

Either because of block faulting, regional warping, or both, the central and southern part of Arizona became an area of aggradation in late Tertiary time, and stream and lake sediments, in places 10,000 feet thick, accumulated in the lower areas. The deposits, though given various names in several local areas, are best known as the Gila conglomerate. Mild volcanic activity accompanied the sedimentation, and lava flows are locally present in and on top of the formation. Relative uplift of the ranges and subsidence of the intermontane trough areas continued intermittently into Quaternary time, and the Gila formation is tilted, faulted, and locally folded. Where uplifted, it is trenched, and the material eroded from it and other sources has been deposited as a relatively thin veneer of Quaternary terrace and stream alluvium.

Tertiary volcanic rocks are nearly everywhere, and in one place or another represent continuing volcanic activity down to the time of the Indians. A résumé of the Tertiary volcanic activity throughout the Tertiary and Quaternary in the mountain and desert regions of Arizona and in the Colorado Plateau from the point of view of age, distribution, and composition is a very inviting study.

In general, the ranges trend northerly in southeastern Arizona and northwesterly in the central and southwestern part. These directions are probably due to the major Late Tertiary faults. Considerable time has elapsed since the last major movements, because extensive pediments have formed across many faults and true fault scarps are few.

CONCLUSIONS REGARDING TECTONIC HISTORY

In southern Arizona sometime during the Triassic and Jurassic, the Paleozoic and Precambrian rocks were folded, intruded by granitic stocks or small batholiths, and deeply eroded. The orogeny is tentatively correlated with the early Nevadan of central and western Nevada and California, and the belt of orogeny is recognized to extend from Arizona to south-central Coahuila in the site of the Late Jurassic Coahuila penin-

sula. It is regarded as a branch of the main orogenic belt of Nevada which continued southward into Mexico west of the Mexican geocyncline.

The Mexican geosyncline transgressed northward in Early Cretaceous time, and considerable thicknesses of Lower Cretaceous clastics were spread over southeastern Arizona. The geanticline on the west was a site of much volcanism, and the volcanic materials contributed to the sediments of the geosyncline. The Concentrator volcanics of south-central Arizona suggest that the volcanic belt continued northwestward into Arizona.

Generally mild deformation followed the Early Cretaecous volcanism and sedimentation, and then over most of the southeastern Arizona Upper Cretaceous strata and volcanics were spread. In Early (?) Tertiary time as part of the Laramide orogeny uplifts with attendant folding and thrusting in the Mountain Region and Sonoran Desert occurred. This was followed immediately by the intrusion of granitic stocks.

Mid-Tertiary time was marked by a varied volcanic activity from place to place, and finally in late Tertiary time block faulting of regional character occurred and caused the widespread deposition of the Gila conglomerate. Some faulting continued in places during the Quaternary, but this was a time chiefly of the development of extensive pediments around the desert ranges.

28.

ROCKIES OF NORTHERN MEXICO

MEXICAN GEOSYNCLINE

The *Tectonic Map of Northern Mexico* by Philip B. King (1947) is reprodued in Fig. 28.1, and on it the belt of Laramide folds can be seen extending southward from New Mexico and Texas into central and eastern Mexico. Preceding the Laramide orogeny and in about the same region a major basin subsided and received a thick complement of sediments. It is known as the Mexican geosyncline. See Plates 10, 11, and 12 for its limits.

Late Jurassic History

In the northern part of the geosyncline in Late Jurassic time, sediments 2600 to 4800 feet thick accumulated, while in the southern part at least 5000 feet of beds were deposited. At the beginning of Late Jurassic time, the subsidence and marine invasion was limited to the southern part, where 2000 feet of dark marine clay, lime mud, and sand were deposited. The southward-lying land was evidently not high, but it was stable and a humid climate prevailed (Imlay, 1943). After this stage, the first widespread marine transgression occurred (Devesian stage), and thick salt and anhydrite beds, associated with red clays, sands, and gravels, were deposited. The salt facies was deposited in northern Central America, southern Mexico, and the southern United States. A thick red-bed facies, at least partly of continental origin, was formed throughout much of northern and eastern Mexico at apparently the same time as the salt facies to the south and north. Both facies transgressed a peneplained surface.

The thickness and extent of the salt layers suggest that the entire Gulf of Mexico was a salt-depositing basin completely enclosed except for a relatively narrow, shallow strait that probably connected with the Atlantic Ocean.

The material composing the red-beds was probably derived from the Central Stable Region on the north, where older red-beds cropped out, and from the geanticlinal areas in western and southern Mexico (Imlay, 1943). Further sinking of the geosyncline brought on normal marine conditions; and lime, clay, and silt were deposited on top of the salt and red-beds. Still later in the late Jurassic time (Kimmeridgian), more red-beds with anhydrite were deposited in the central parts of the geosyncline.

At this stage in late Jurassic time an uplift formed the Coahuila peninsula extending southward across western Coahuila and eastern Chihuahua as far as the Parras basin of southern Coahuila (Kellum, 1944). See map of Late Cretaceous, Plate 12. Coarse clastics that were deposited marginal to it on three sides during latest Jurassic as well as Early Cretaceous time attest a fairly high topography to the peninsula.

In the description of the Coahuila system, an orogenic belt was described, part of which, at least, occupied the position of the later Coahuila peninsula. It appears that the folded and faulted Permian beds were first intruded, then eroded, and then epeirogenically uplifted before the Upper Jurassic beds were deposited in the area. See Plate 13. The belt of volcanoes on the west that supplied much of the Permian sedi-

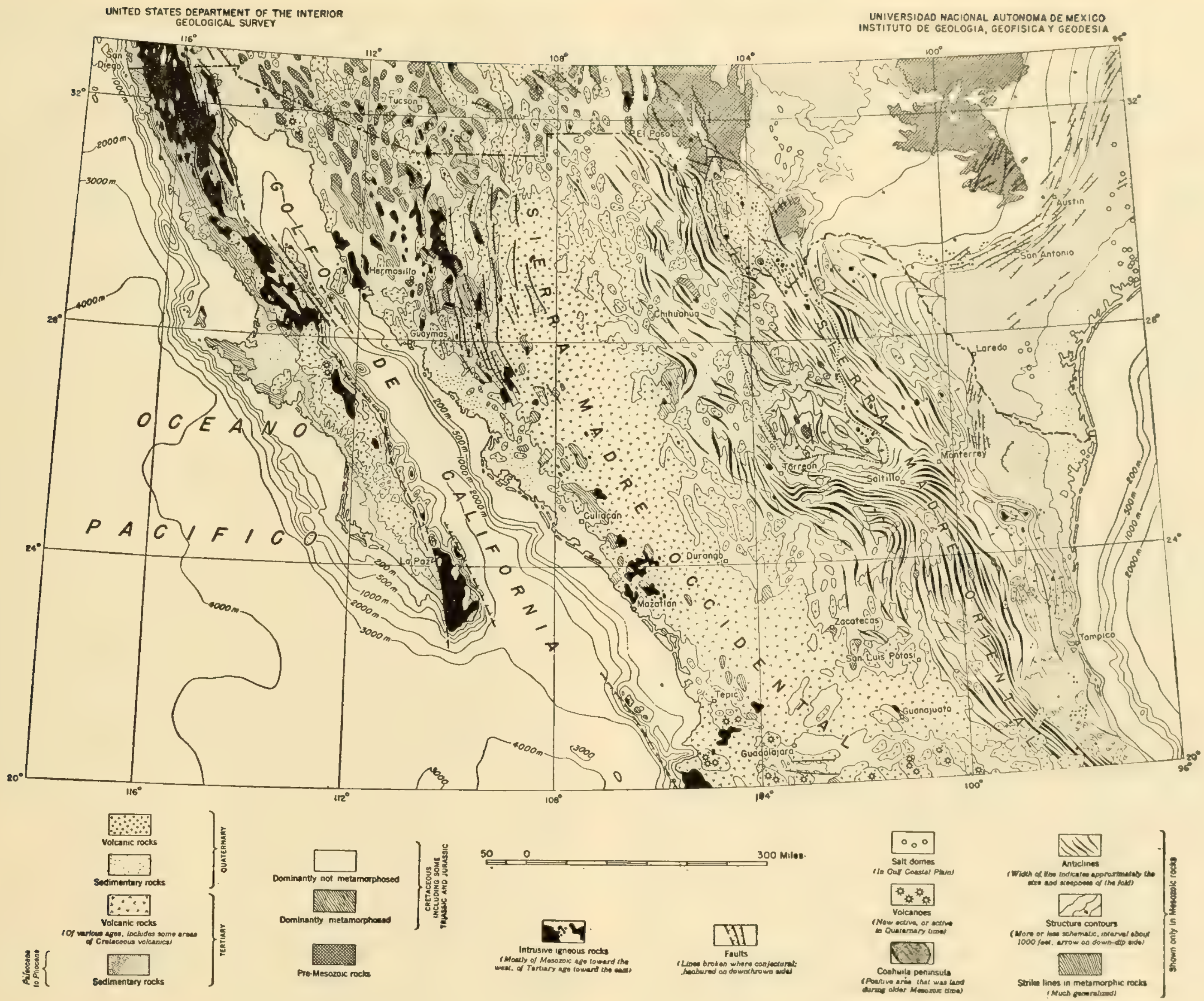


Fig. 28.1. Tectonic map of northern Mexico. Reproduced from P. B. King, 1942b.

ments sank, and the area became the Mexican Upper Jurassic and Cretaceous geosyncline.

Northeast of the Coahuila peninsula, there is evidence of a narrow promontory, called the Oriental geanticline by Imlay. The structure for the most part was barely emergent.

Early Cretaceous History

During Early Cretaceous time, the Mexican geosyncline sank over 12,000 feet and received its greatest load of sediments. The contours of the tectonic map, Plate 11, were drawn on the basis of thicknesses given by Imlay (1944), and in conformity with the Coahuila peninsula and the Orental geanticline as outlined by Kellum and Imlay. The Coahuila peninsula was overlapped considerably, and by Aptian time (middle Early Cretaceous) it was completely submerged and for the rest of Cretaceous time was a platform on which 1500 feet of lagunal deposits accumlated.

The land along the western margin of the Mexican geosyncline, at least in the northern part, had probably suffered intense deformation during the late Early Cretaceous. Over 10,000 feet of beds of Aptian age lie along its eastern margin in central Sonora. The outcrops farthest west consist of andesite flows, tuffs, and agglomerates, but this volcanic facies is replaced to the east by a marine limestone, shale, and sandstone facies. According to R. E. King (1939), there is evidence of great oscillations in level of the sea, with repetition of cycles of marine and continental deposits.

In northern Sonora, thick coarse conglomerates of the same age as the volcanics of central Sonora occur. They increase in thickness southeastward from Bisbee, Arizona, and are at least 5000 feet thick only 30 miles to the south. The boulders are in part large and angular and, together with the large volume, show that the sea was bordered by steep shores, and that the southwestern landmass was suffering active deformation. The deposition and causative orogeny was rapid, because the time represented by the conglomerates and related sediments is but a fraction of Cretaceous time (Imlay, 1939). The presence of finer clastic sediments, as well as coal, higher in the section indicates times of lowered lands, broader littoral zones, and marginal swamps. The lithographic character of the Bisbee group shows that the landmass to the south was composed of Precambrian gneisses and schists, and Paleozoic quartzites and limestones similar to those outcropping at present in the Bisbee district and locally in northern Sonora (Imlay, 1939).

Still farther west than Sonora in Baja California, Lower Cretaceous rocks have been identified, and their nature is significant regarding the belt of orogeny west of the Mexican geosyncline. They occur along the west shore of the northern part of the peninsula and consist of conglomerates, quartzites, tuffs, and agglomerates, with thick lava flows interbedded. They are cut by dikes and large stocks. In some localities, the intrusive rocks predominate over the sediments and pyroclastics, and in places there is much metamorphism. Unaltered, or but little altered, sandstones and shales appear in places, and limestone also occurs. The metamorphosed Cretaceous rocks may be equivalent to schists and coarse, massive, white granite that are widely distributed southward down the peninsula. The granitoid rocks that intruded the Lower Cretaceous beds are probably early Late Cretaceous in age. Chapter 30 is devoted to the geology of Baja California and Sonora, and should be consulted for further details.

The composition of the Lower Cretaceous rocks is that of the volcanic archipelago type, and indicates an associated orogenic belt. They appear to be separated from the deposits of the Mexican geosyncline by the Occidental geanticline, but not enough is known of the distribution and lithologic variations of the Lower Cretaceous strata in this region to demonstrate the interpretations. Refer further to Chapter 30.

Late Cretaceous History

Upper Cretaceous deposits are widely distributed from the Santa Ana Mountains southeast of Los Angeles (Woodford, 1939) throughout the length of Baja California to Todos Santos. They are separated from the strata of Early Cretaceous age by a period of intrusion and metamorphism, and were themselves faulted and in places folded before the Cenozoic sediments were laid down. For further details, see Chapter 30.

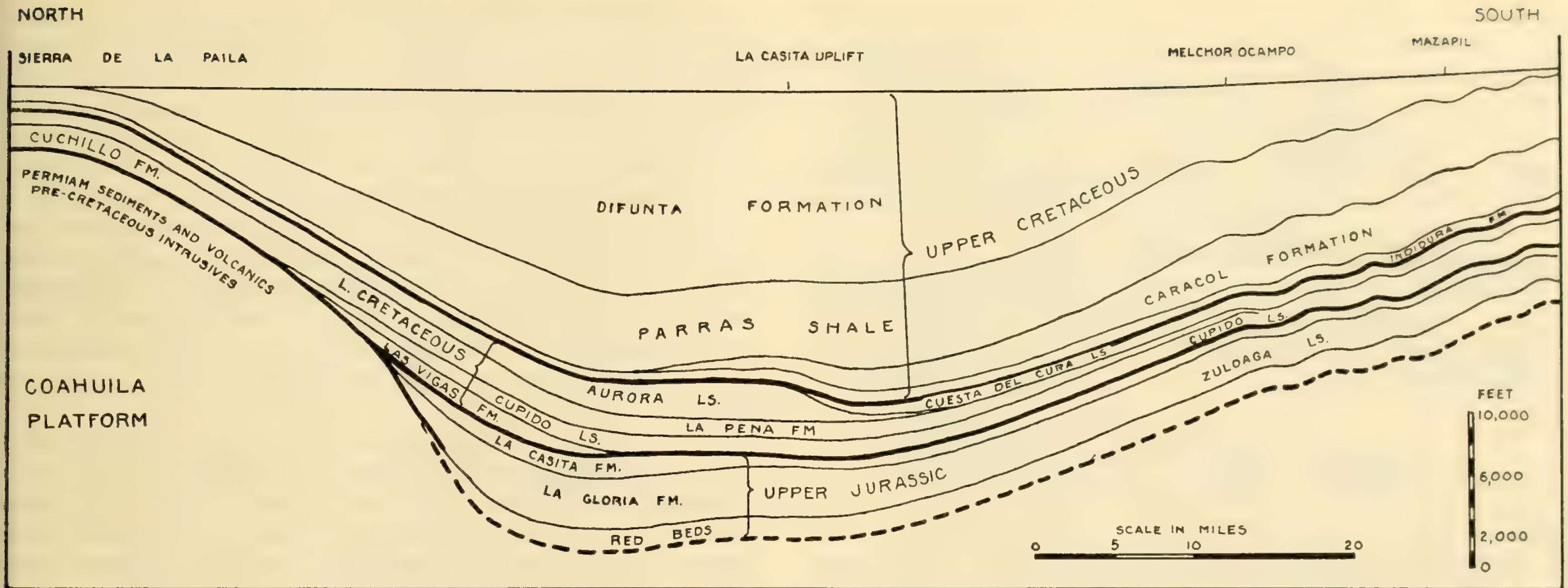


Fig. 28.2. Parras trough at close of Cretaceous sedimentation. After Imlay, 1939.

The orogenic belt of western Sonora continued active during Late Cretaceous time and crowded the Mexican geosyncline eastward. At two different times, the land rose sharply, and thousands of feet of sediment were deposited along the western edge of the trough (now in eastern Chihuahua); first shale and sandstone, and later conglomerate. In northern Zacatecas, southeastward from Sonora but still along the western margin of the geosyncline, thousands of feet of tuffaceous beds were deposited.

An east-west trough subsided over 15,000 feet in southern California. Figure 28.2 shows the sediments in it and the Coahuila platform to the north. The depression has been named the Parras trough from the present Parras basin in which the Upper Cretaceous sequence crops out (Imlay, 1944).

SONORAN REGION

Very little is known about the Paleozoic history of Mexico. The Permian of the Mexican state of Coahuila has been treated in Chapter 14. On the west side of the country in Sonora, additional Permian beds have been noted (R. E. King, 1939). They consist of limestones with abundant crinoid stems and fusulinids. Reefs of massive limestone about 1500 feet thick grade laterally into lesser thicknesses of well-bedded, darker limestone. Permian strata may have occurred, originally, east of the central Sonoran outcrops, because cobbles of fossiliferous Permian limestone are found in the basal conglomerate of the Cretaceous there (R. E. King, 1939). Present data are not sufficient to outline the basin in northwestern Mexico, and the tectonic significance of the Sonoran exposures is, there-

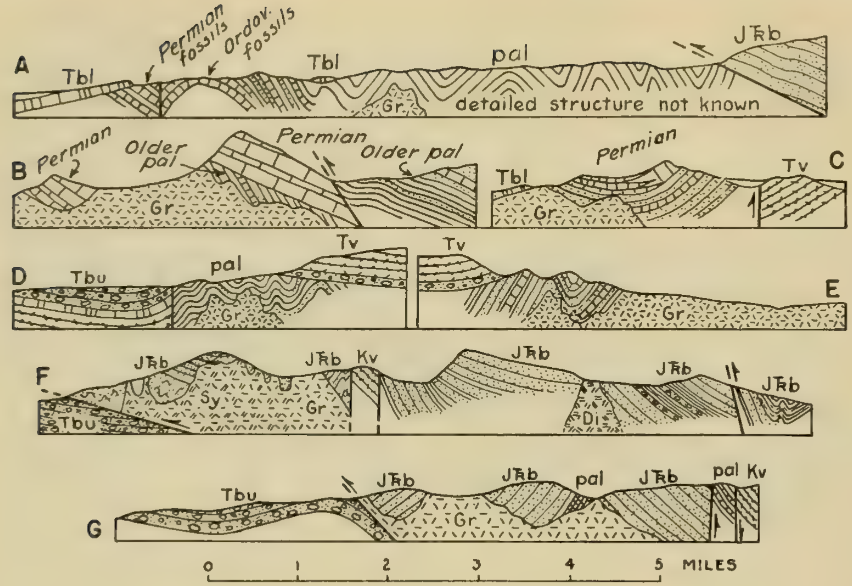


Fig. 28.3. Sections illustrating structures and stratigraphy in central Sonora, Mexico. A, Clasita area; B, Urro Cobachi area; C, Zubiate area; D, El Trigo area; E, Arroyo Arenosa area; F, Southern Sierra de San Javier; G, vicinity of Guamochil. pal, Paleozoic rocks; JTRb, Barranca formation; Kv, Cretaceous volcanics; Tv, early Tertiary volcanic rocks; Tbl, lower member of Báucarit formation; Tbu, upper member of Báucarit formation; Gr, granite; Di, diorite; Sy, syenite. After R. E. King, 1939.

fore, obscure. Not helping to clarify the obscurity is the average eastward strike of the younger rocks. See map, Plate 8. Late Permian or early Triassic folding is indicated by generally greater metamorphism and folding in the Permian than in nearby Upper Triassic and Cretaceous rocks, as well as by the divergent strikes. The immediate impulse is to relate the Sonoran east-west trends to the Coahuila, but then the belt of volcanics that supplied much material to the Coahuila Permian basin apparently lay between, and it would therefore seem that the two did not form a single continuous tectonic system. The peculiar thing is that the volcanic archipelago type of sediments is on the east in this region and the inland basin type on the west, just the opposite from the distribution to the north in the western United States and Canada. Because the Sonoran Permian is of the inland basin type, it might be supposed that the Pacific orogenic belt lay considerably west. Since the peninsula of Baja California is made up in large part of Cretaceous batholithic intrusions, evidently a continuation of the great Sierra Nevada batholithic belt, and since the batholithic belt coincides strikingly with the Permian trough in the United States, British Columbia, and southeastern Alaska, it can also be supposed that the Permian orogenic belt paralleled the peninsula and perhaps in part lay west of it. It is evident that this is supposition, but possibly a reasonable guess in the absence of factual information. The interpretation rendered on the Permian tectonic map, Plate 8, does not produce a meaningful tectonic pattern and is probably not correct, but various other arrangements seemed even less tenable. We must await more field work in western Mexico.

Paleozoic rocks older than Permian are rare in Mexico. In central Sonora, Ordovician limestone, sandstone, and conglomerate have been identified (R. E. King, 1939). See cross sections, Fig. 28.3. In southernmost California, an outcrop of marble is Mississippian. Farther north in the San Bernardino Mountains, the Furnace limestone may be Mississippian (Woodford, 1939). It is bounded above and below by quartzite formations. Very similar massive dolomitic limestones and somewhat similar quartzites are widely distributed in the Perris and San Gabriel mountains. The Arrastre quartzite of the San Bernardino Mountains underlies the Furnace limestone, and is so far below the fossiliferous horizon that it is probably pre-Mississippian (Woodford, 1939).

The earliest formation of the Mesozoic in Mexico is the Barranca. It is Late Triassic and Early Jurassic in age, and crops out extensively in northwestern and central Sonora (R. E. King, 1939). There are also isolated exposures in southeastern Sonora and western Chihuahua. In the ranges bordering the Rio Yaqui, the formation is wholly of continental origin, with a thickness of more than 3300 feet. Three members are recognizable. The upper and lower divisions are massive sandstones with some interbedded dark shale. The middle member consists of shale and thin-bedded sandstone with layers of coal and graphite. Farther east a short distance, the formation is 4250 feet thick. In northwestern Sonora, a clastic

section is about 7350 feet thick. The lower part is Late Triassic, and the upper part is Early Jurassic.

EL PASO–RIO GRANDE THRUST BELT

When the Late Cretaceous seas of the Mexican geosyncline finally withdrew from the El Paso–Rio Grande area, the sedimentary veneer on the Precambrian crystallines was of appreciable, although variable, thickness. In the northern Quitman Mountains, it was 10,000 feet thick. The ancestral Diablo Range along the north side of the Rio Grande was an area of thinning of Pennsylvanian and Permian strata (see the paleotectonic maps of Plates 7, and 8), and the Coahuila platform to the south and west was an area of marked thinning of the Lower Cretaceous (see paleotectonic map, Plate 15).

The Laramide compressional forces then gripped the El Paso–Rio Grande area and subjected it to intense deformation. Examine the map of Fig. 28.1. The rocks were highly folded and thrust-faulted (Huffington, 1943). Three thrusts are prominent, namely, the Devil Ridge, Red Hills, and Quitman. See cross section Y of Fig. 25.16. They occur in the Quitman and Malone Mountains and in Devil Ridge and, altogether, make a zone about 75 miles long from the Hueco basin on the north to the Chinati Mountains on the south. Folds and thrusts are known in a broad belt west of the Quitman Mountains in Chihuahua, Mexico. The areas of bedrock are few in northern Chihuahua, and these are little known geologically. Consequently, the western limit of the Laramide Sonoran Rockies there is indefinite. They probably merged with the Sierra Madre Occidental Rockies to form a great broad belt of deformation. Much of this area is now in the Basin and Range structural province because of the superposition of later block faults on the Laramide folds and thrusts.

The El Paso–Rio Grande thrust belt probably extends far enough southeastward to merge with the Sierra Madre Oriental system of Laramide ranges in Coahuila. Its eastern boundary is sharply defined between the Malone Mountains and the Sierra Blanca. Folding and thrusting are prominent in the Malone Mountains, whereas the strata in the Sierra Blanca are but slightly folded and cut by a few small normal faults.

Igneous activity was conspicuous in the thrust belt, principally in the Quitman Mountains. The same igneous province spread over an extensive part of the domes and basins of the foothill province, viz., the Davis Mountains volcanic field, the Chisos Mountains, and the Serranias de Burro uplift. See the *Tectonic Map of the United States.* In the Quitman Mountains, the folding and thrusting are followed by extensive erosion and then eruption of a volcanic series of rhyolites, trachytes, and andesites. The volcanic rocks sagged and were intruded by a ring dike of diorite. Then an intrusion of quartz diorite followed, and afterward the quartz monzonite Quitman pluton (Huffington, 1943). This activity, if related to the Davis Mountains volcanics, occurred in Eocene and Oligocene time.

PLATEAU CENTRAL AND SIERRA MADRE ORIENTAL

The Jurassic and Cretaceous geosyncline of Mexico and the Coahuila peninsula have already been described. Refer particularly to the paleotectonic maps of this book. During the Laramide orogeny, the thick sediments of the geosyncline were caught in compressive forces and severely folded, but the thinly veneered peninsula was only slightly deformed. In places, later block faults are superimposed on the Laramide folds; but for the most part, the present-day mountains of the Sierra Madre Oriental and the high surface of the Plateau Central are the result of a long chronicle of erosion and alluviation. The two provinces from the eastern half of the highlands of northern Mexico. They are closely related tectonically but are somewhat different in surface features. The Plateau Central is an area of relatively low relief, but high altitude, and consists of wide bolson plains from which rise mountains composed largely of the folded sedimentary rocks. The Sierra Madre Oriental is a region of high relief along the east side of the Plateau Central and is composed of parallel mountain ranges also of folded sedimentary rocks.

Figure 28.4 is a typical example of the folds in the geosynclinal parts of the province. They are tight to the point of being isoclinal, overturned, and even fan-shaped in places, and closely packed; but in spite of the

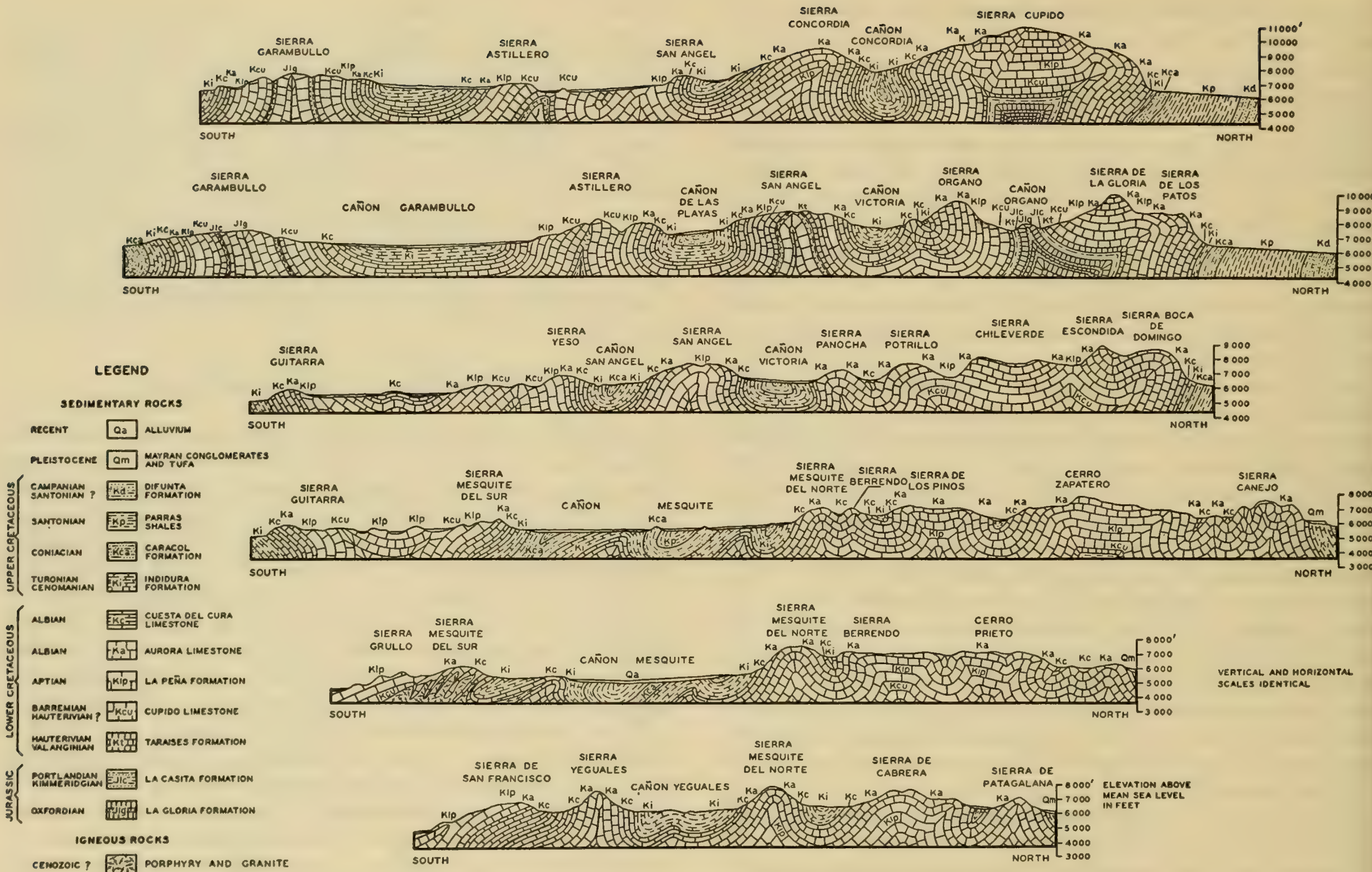


Fig. 28.4. Cross sections of the middle part of the Sierra de Parras, Coahuila, Mexico. Reproduced from Imlay, 1937.

intense shortening, few thrust faults developed. The Sierra Madre Oriental, west and northwest of Tampico for a distance of about 100 miles, is made up of several ranges, the Sierra Cucharras, the Sierra Tanchipa, and the Sierra del Abra. These also consist mostly of folds, but at a number of places Kellum (1930) has interpreted a thrust structure along their east front. His drawings are reproduced in Fig. 28.5. This belt of thrusting is approximately in line with the El Paso–Rio Grande thrust belt farther north. Several large thrusts along the west side of the Sabinas basin seem to connect the northern and southern thrusts and to form a belt about 800 miles long. See *Tectonic Map of the United States* and Fig. 28.1.

The region formerly occupied by the Coahuila peninsula is one, according to Kellum *et al.* (1936a):

> . . . of broad, gentle folding and includes the great brachyanticlines or periclinal folds of the Sierra de la Paila, the Sierra de los Alamitos, and the Sierra de García, in the east. It also includes the Sierra del Venado, the Sierra del Sobaco, the Sierra del Tlahualilo, the Sierra de Campana, and related ranges, in the west. Undoubtedly, it takes in many mountain ranges lying to the north of the western group, but these have not been studied in sufficient detail to demonstrate the regional structural plan. The eastern group of ranges also has never been studied in detail, but the general structure, as seen from the south and as reported by Böse, Kane, and others who have crossed them, is a gentle uplift. The western group is essentially the same but differs in that erosion has progressed much further and divided the broad, gentle uplift into numerous ranges, more or less separated by valleys filled with alluvium.
>
> These ranges are composed, in large part, of the gypsum facies in the Cuchillo formation. This is an easily eroded unit, and, where the gypsum and marl predominate in the section, the mountains have been cut down more rapidly than where limestone predominates.
>
> The structure of the Cretaceous rocks in the mountains bordering the valleys of Las Delicias and Acatita illustrates the type of folding characteristic of the central province (Coahuila Peninsula). The major structure between the two valleys is a broad, composite, anticlinal uplift, trending northwest-southeast and plunging in both directions. Superimposed upon it are many sharp, persistent folds, parallel to the central axis. Minor cross folds, ordinarily nonpersistent and with gentle dips, appear to reflect topographic irregularities in the basement rocks. The main axis of the major anticlinorium extends along the western margin of the range, in its northweastern part, but to the southeast the axis crosses the central part of the mountain area. Limited observation on the minor anticlines southwest of this axis indicates that they tend to be asymmetrical, with the steeper dip on the southwest. The major structure of the Cretaceous rocks in the Sierra del Tlahualilo, west of the Acatita Valley, is a broad, gentle fold, almost perpendicular to this major trend, cross the range—one, at its north end; the other, about 15 miles farther south. These are believed to reflect topographic features, or zones of displacement, in the underlying basement rocks.

Figure 28.6 shows an example of the structure that developed over the site of the former Coahuila peninsula.

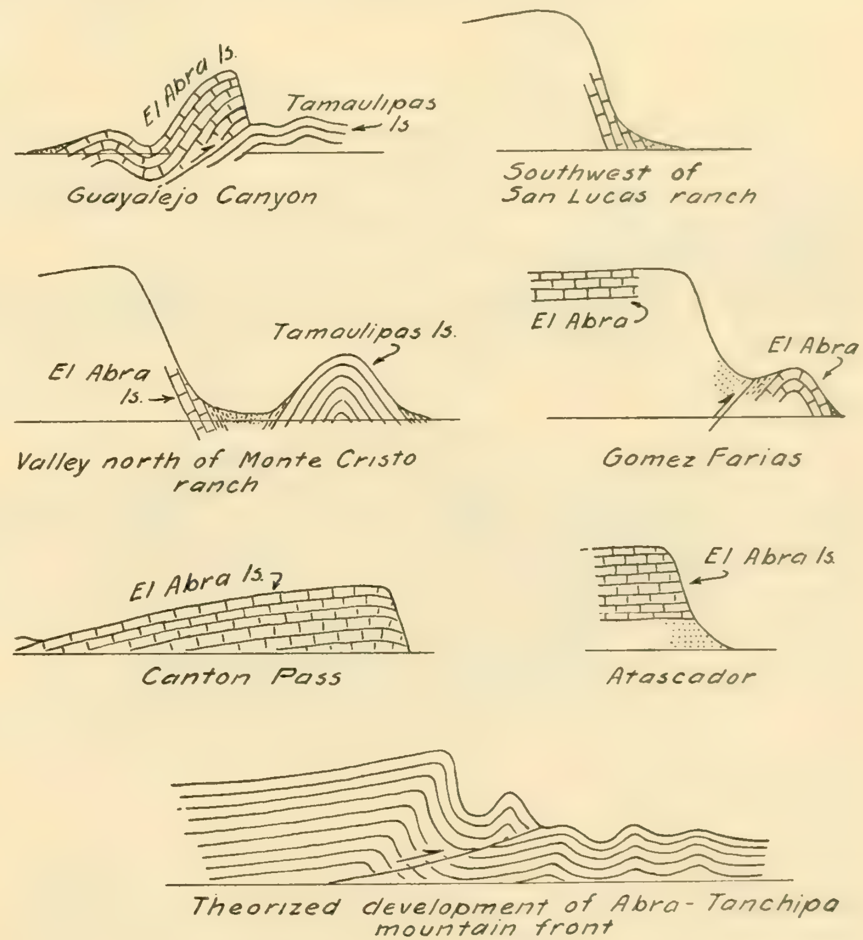


Fig. 28.5. Sierra Madre Oriental front west of Tampico. After Kellum, 1930.

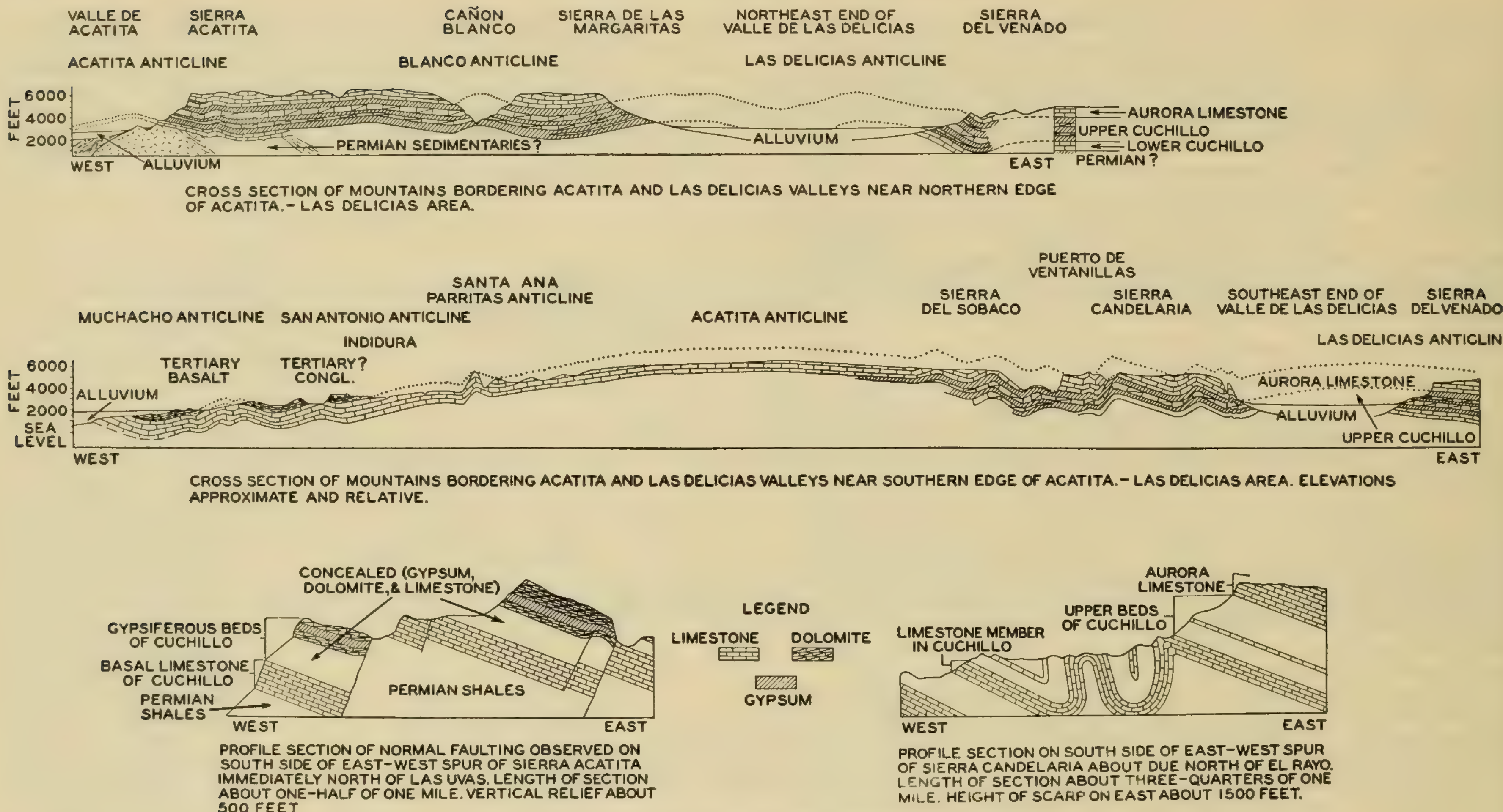


Fig. 28.6. Cross sections of the Acatita and Las Delicias area, Coahuila, Mexico. Reproduced from Kelly, 1936.

PARRAS SYNCLINORIUM

South of the Coahuila peninsula is a belt of sinuous folds that trends approximately east-west. It is about 130 miles long, and tapers from a width of 40 miles in the west to 20 miles in the east. The folds developed out of the sediments of the Parras trough (see paleotectonic map of the Late Cretaceous). The anticlines plunge to the west beneath the alluvial plain of the Laguna de Mayrán; they are steepest on their north flank, and are usually overturned. Along any one occur domes and saddles, of which the domes are more overturned (de Cserna, 1956).

At the western end of the Parras basin, the Sierra de Hispaña overthrust occurs. See Fig. 28.7. The thrust sheet has ridden northeastward

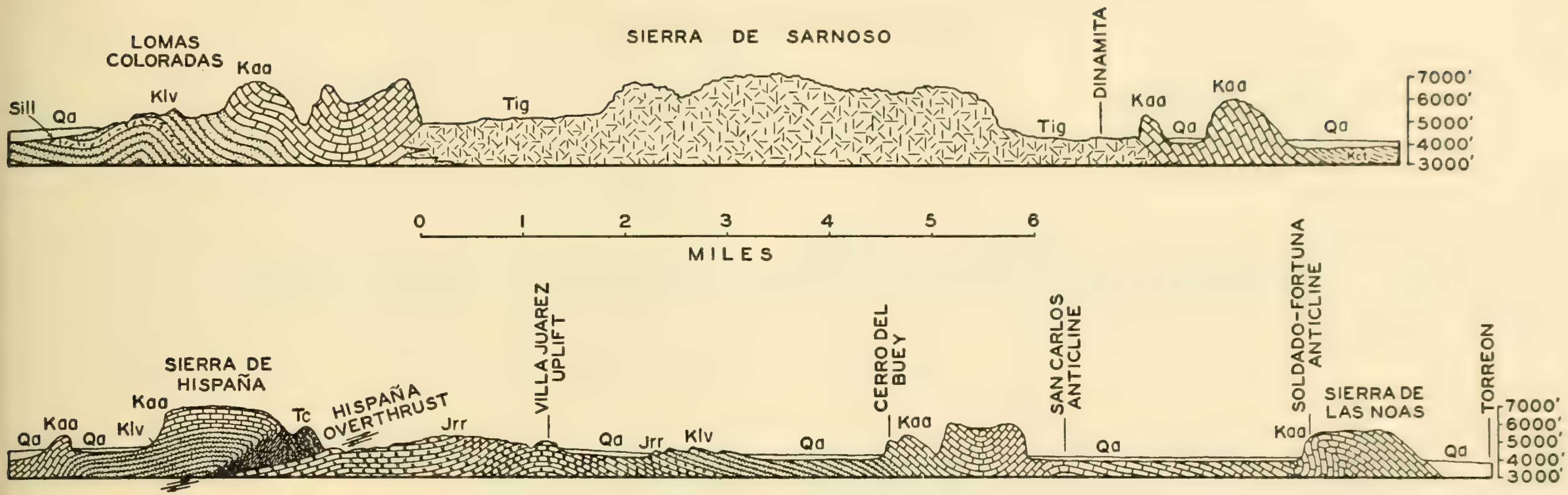


Fig. 28.7. Cross sections of the mountains west of the Laguna district, after Kellum, 1936. Tc, Late Cretaceous or Early Tertiary conglomerates; Kct, Cenomanian-Turonian sh. and ls.; Kaa, Aptian-Albian ls.; Klv, Torcer-Las Vigas series; Jrr, Red Rock series; Tig, igneous intrusives.

against the buttressing and less deformed Coahuila peninsula. To the west of the Sierra de Hispaña thrust, a high mountainous mass consists of tight folds overturned toward the northeast. This zone has not been traced in ranges to the northwest, but it undoubtedly continues in that direction.

The east-west belt of folding veers south-southwesterly at Saltillo and merges with and forms the Sierra Madre Oriental. Quoting from Kellum *et al.* (1936a) again:

In this region, intensive compression has developed a series of overturned or fan-shaped anticlines and synclines in Cretaceous and Jurassic rocks, with enormous horizontal shortening. The axes of these folds trend, in general, east-west and pass eastward by a rather short curve to a southeast direction. At the west end, where they close and plunge into the Parras Basin, they become symmetrical, and then are overturned westward, parallel to the strike of their axes. No important faulting has been recognized.

East of the Sierra Madre Oriental, this prominent zone of east-west folds is present in central Tamaulipas in the San Carlos Mountains, which rise out of the coastal plain about midway between the Cordilleran front and the Gulf Coast. The San Carlos Mountains are a broad, arcuate geanticline [arch as defined in this book], trending in a general easterly direction, with the convex side to the south. Superimposed upon this major structure are numerous, low flexures, parallel to it, and also a number of domes produced by igneous intrusions. The axes of folds in this geanticline, are not the continuation of axes in the Sierra Madre Oriental, but are the continuation of axes which lie east of, and parallel to, the Cordilleran front, farther northwest, and are turned eastward in the zone of cross-folding.

West of the Sierra Madre Oriental and south of the Parras Basin the structure of this belt has been studied in a number of areas. In the region of Mazapil–Concepción del Oro, in northern Zacatecas, the mountain ranges are anticlinal and trend generally eastward; their structure is complicated by several faults and by the presence of intrusive masses. The Sierra de Santa Rosa and the core of the Sierra de la Canutillo show a slight tendency toward fan structure. The Concepción del Oro anticline, a southeast continuation of the Sierra de la Caja structure, is overturned toward the northeast and is crossed by a fault.

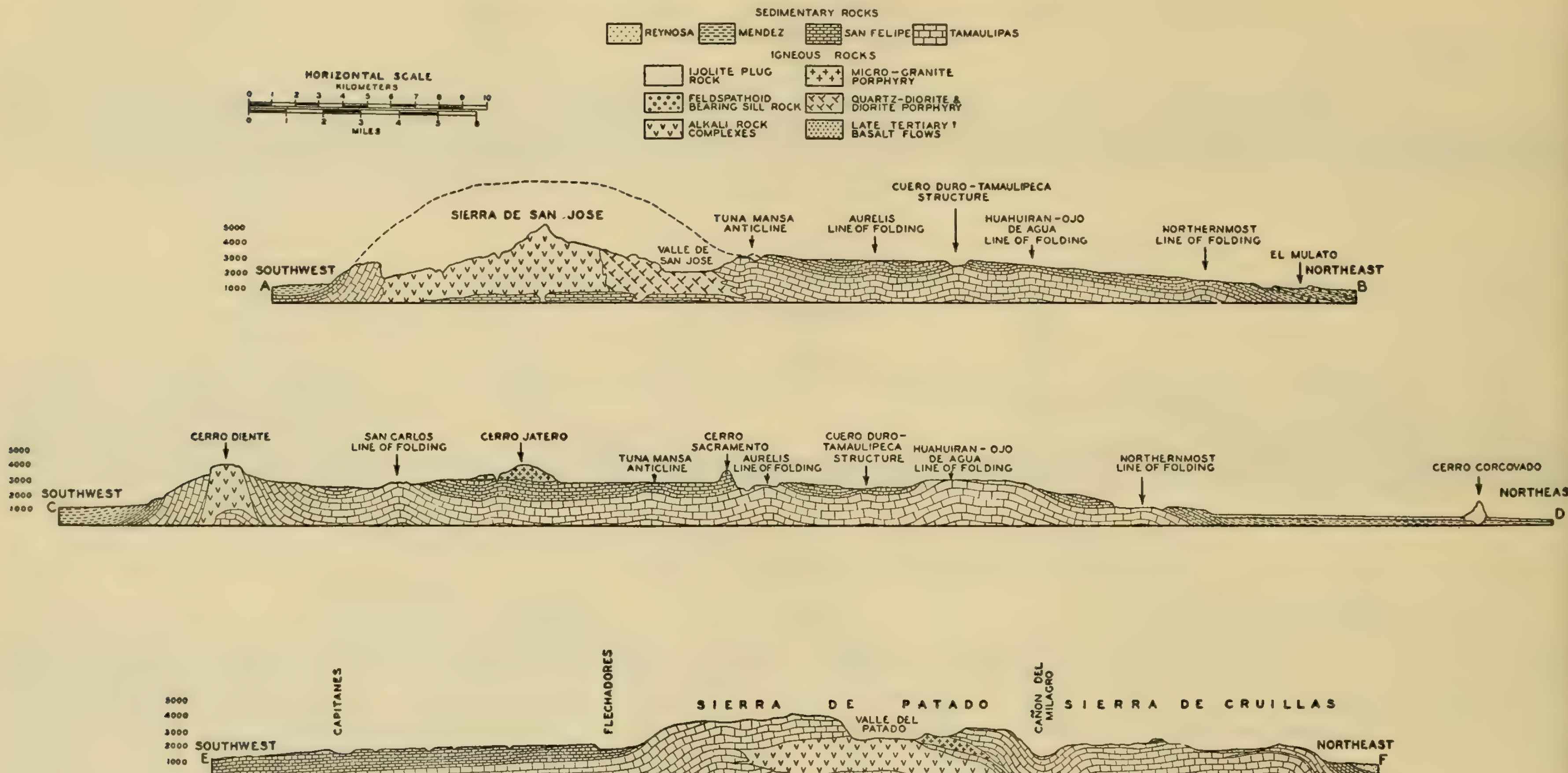


Fig. 28.8. Cross sections of the San Carlos Mountains, Tamaulipas, Mexico. Reproduced from Kellum, 1937.

OROGENIC HISTORY

In review of the orogenic history of northeastern Mexico, it has been concluded that the late Paleozoic belt of folding and thrusting of the Marathons of western Texas curved southward and probably extended into the site of and formed the later Coahuila peninsula. From the southern part of the peninsula, the late Paleozoic orogen turned eastward to Monterey, according to Humphrey (1947) and then south-southwesterly for a long, but unknown, distance. It seems doubtful that the orogenic belt could have endured throughout more than early Mesozoic time, when epeirogenic movements may have rejuvenated segments of it which served as barriers to marine invasions from the Gulf region. The Mexican geosyncline developed in Jurassic and Cretaceous time on the western and southern (foreland) sides of the Paleozoic belt, but sedimentation

also occurred on the eastern (hinterland) side. In Late Jurassic time, movements of the southern part of the peninsula adjacent to the geosyncline are reflected in the sediments; and these in turn reflect the Nevadan orogeny. In the Early Cretaceous, minor and local movements in the Coahuila peninsula furnished some coarse sediments to the seas. The peninsula was finally submerged in Aptian time but continued to act as a relatively high and stable mass. In the Late Cretaceous, considerable thicknesses of sediments were deposited along the southern border of the peninsula in the Parras trough, whose position coincides with the margin of the Late Jurassic seas.

In the Early Tertiary, the deposits adjacent to the peninsula and its eastward and southward extension were deformed into long narrow folds by forces acting about normal to the western and southern border of the buttressing mass, which itself was only slightly deformed. Along the east side another belt of folds was formed. This belt is overthrust eastward along its east margin, and a foothill belt of structures was formed in front.

FOOTHILL BELT

In places along the inner margin of the Gulf Coastal Plain and not far east of the east front of the Sierra Madre Oriental are several low ranges. About 60 miles southwest of Del Rio on the Rio Grande is the Serranias del Burro; to the southeast of this is the Sierra Lampazos; still farther to the southeast is the Sierra de San Carlos; and then northwest of Tampico is the Sierra Tamaulipas. These are generally broad folds and domes that rise from the nearly horizontal beds of the Coastal Plain and expose gently arched Lower Cretaceous limestones along their crests. Some of these mountains, notably the Sierra de San Carlos, contain igneous plugs and laccoliths (Fig. 28.8) in part of alkalic composition. The San Carlos Mountains also contain gentle, east-west trending folds superposed on the dome; and these, Kellum (1936, 1937) believes, reflect the eastward trend of the Parras basin folds. The folds, in fact, are found, although ill-defined, still farther east in the Sierra de Cruillas in central Tamaulipas, where they have a northeast trend, and there disappear under the Coastal Plain.

About 60 miles west of Tampico is the Sierra del Abra, which there forms the east front of the Sierra Madre Oriental. It is an uplift in part of eastwardly overturned Middle and Upper Cretaceous beds (Kellum, 1930). East of the Sierra del Abra is a Cretaceous and Tertiary basin and then the buried Tamasopo ridge about halfway between the Sierra Madre and the coast. Kellum believes that this buried ridge has had a history similar to the Sierra del Abra. It is the site of the remarkable "southern" oil fields of Tampico-Tuxpam district of Mexico. Basic intrusions and extrusions are present in both the Tamasopo buried ridge and the Sierra del Abra.

29.

COAST RANGES OF THE PACIFIC AND THE SAN ANDREAS FAULT SYSTEM

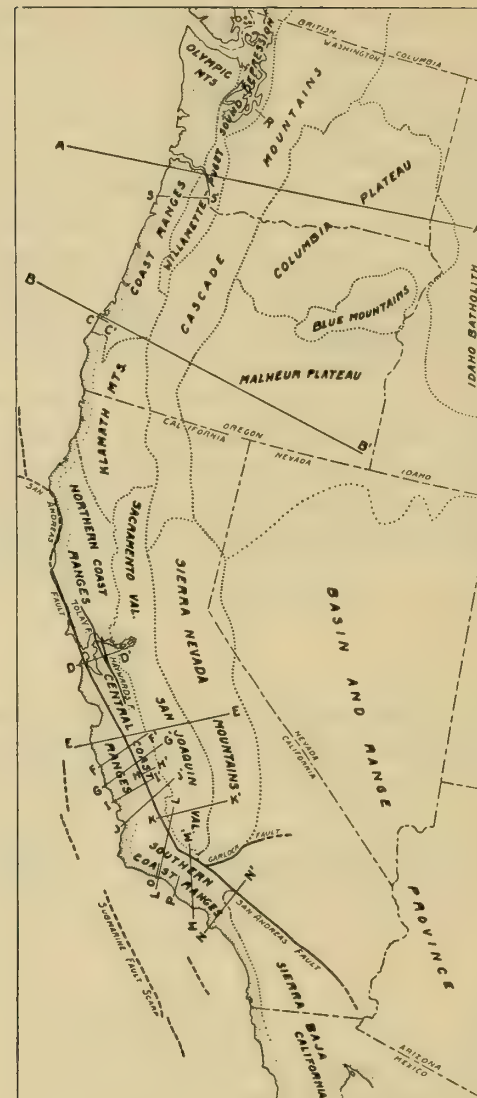


Fig. 29.1. Index map of the Coast Ranges and associated geological provinces of Washington, Oregon, and California, showing the lines of cross sections and the San Andreas fault and possible associates.

MAJOR DIVISIONS

A belt of mountains parallels the present coast in Washington, Oregon, and California and, except as noted, the mountains are known as the Coast Ranges. They are separated from more interior chains by broad depressions, in Washington and Oregon called the Willamette–Puget Sound depression, and in California called the Great Valley. The index map, Fig. 29.1 shows these features. The Coast Ranges have had a prolonged Tertiary and Quaternary history, and their development occurred parallel in time with the Laramide Rockies and the ranges and valleys of the Great Basin.

The division of the Coast Ranges from San Francisco Bay southward to Santa Maria will here be designated the Central Coast Ranges (Fig. 29.2), and the division north of San Francisco Bay to the Klamath Mountains will be called the Northern Coast Ranges. A division in southern California with pronounced east-west trends including the Santa Barbara, Ventura, and Los Angeles districts is referred to as the Southern Coast Ranges or Transverse Ranges. The Coast Ranges of Oregon and Washington are a unit geologically and will be considered as a fourth division. They are separated from the Northern Coast Ranges of California by the Klamath Mountains, which are part of the Nevadan orogenic belt.

The Sierra Baja California of southermost California or the Peninsular ranges, and the peninsula of Baja California is a fifth division, but will be discussed in Chapter 30. It is a complex of Nevadan geology and later Cretaceous and Tertiary beds affected by folding and block faulting.

Still another division, the sixth, remains to be mentioned, namely, the submarine area south of the Transverse Ranges. This ocean bottom has been found in recent years to be one of rugged relief, and the researchers who have ventured a diagnosis of the topography there agree that it is part of the continental framework. It is discussed in Chapter 32.

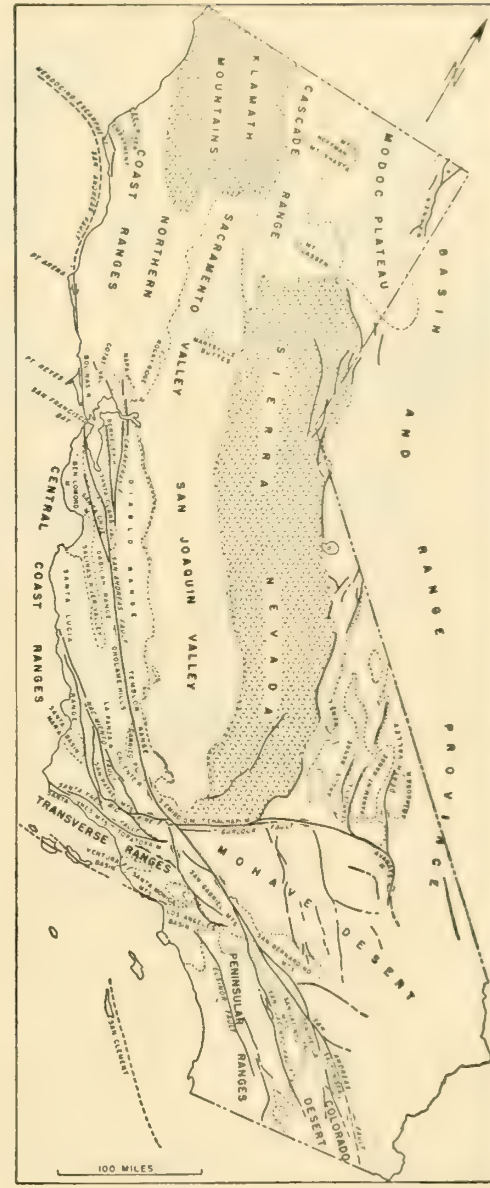


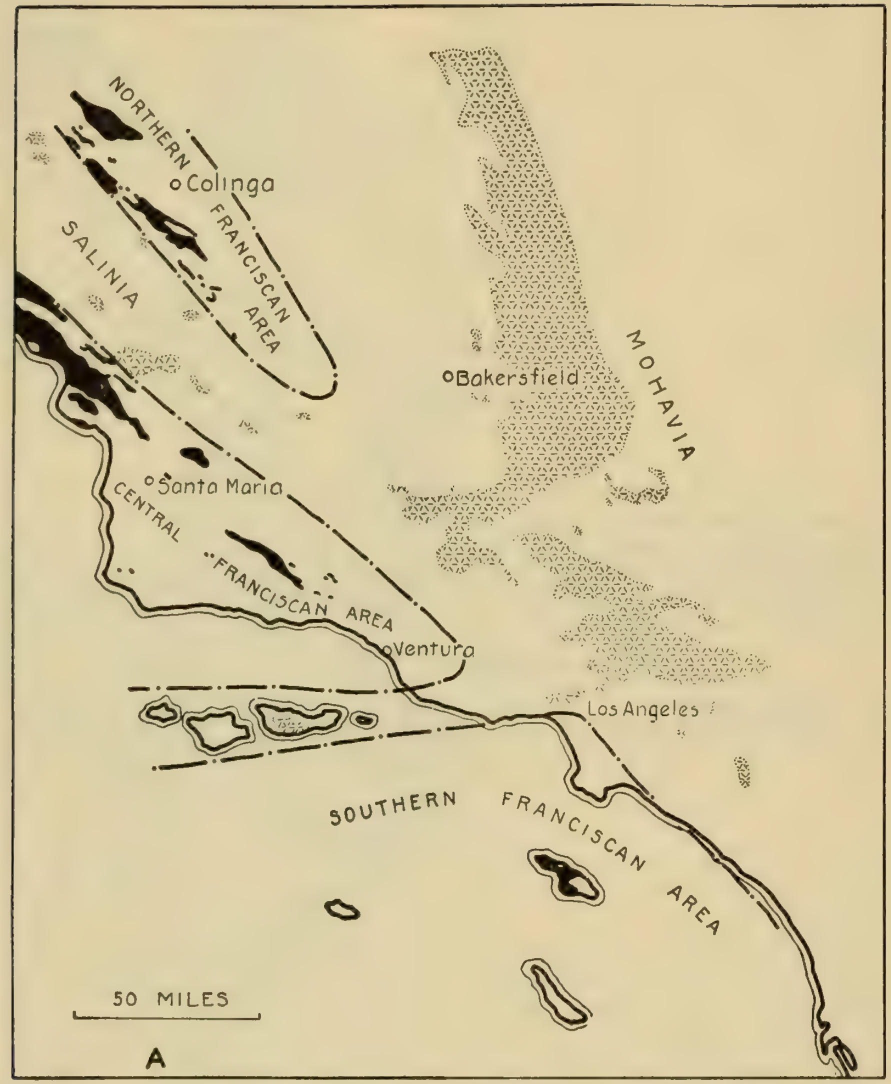
Fig. 29.2. Index map of the Coast Ranges and fault systems of California. Compiled from the fault map of California (1955), from Dibblee, unpublished maps, and other sources.

CENTRAL COAST RANGES OF CALIFORNIA

San Joaquin Embayment and the Diablo Uplift

In the tectonic map of the Late Cretaceous, Plate 12, it will be seen that the uplift of Salinia separated a basin of sedimentation to the north and south in the region of the Central Coast Ranges. The chief change that occurred in Early Tertiary time is that Salinia altered position and size somewhat and became the Diablo uplift; and another small uplift, the San Rafael, came into existence just to the south. The details of these changes are shown in the paleotectonic maps of Fig. 29.3. Also, the trough of deposition, the San Joaquin embayment, became less constricted opposite the uplifts and received from 5000 to 15,000 feet of sediments in the site of the Central Coast Ranges, and 30,000 feet in the Southern Coast Ranges.

Under a later heading, the San Andreas fault system, it will be shown



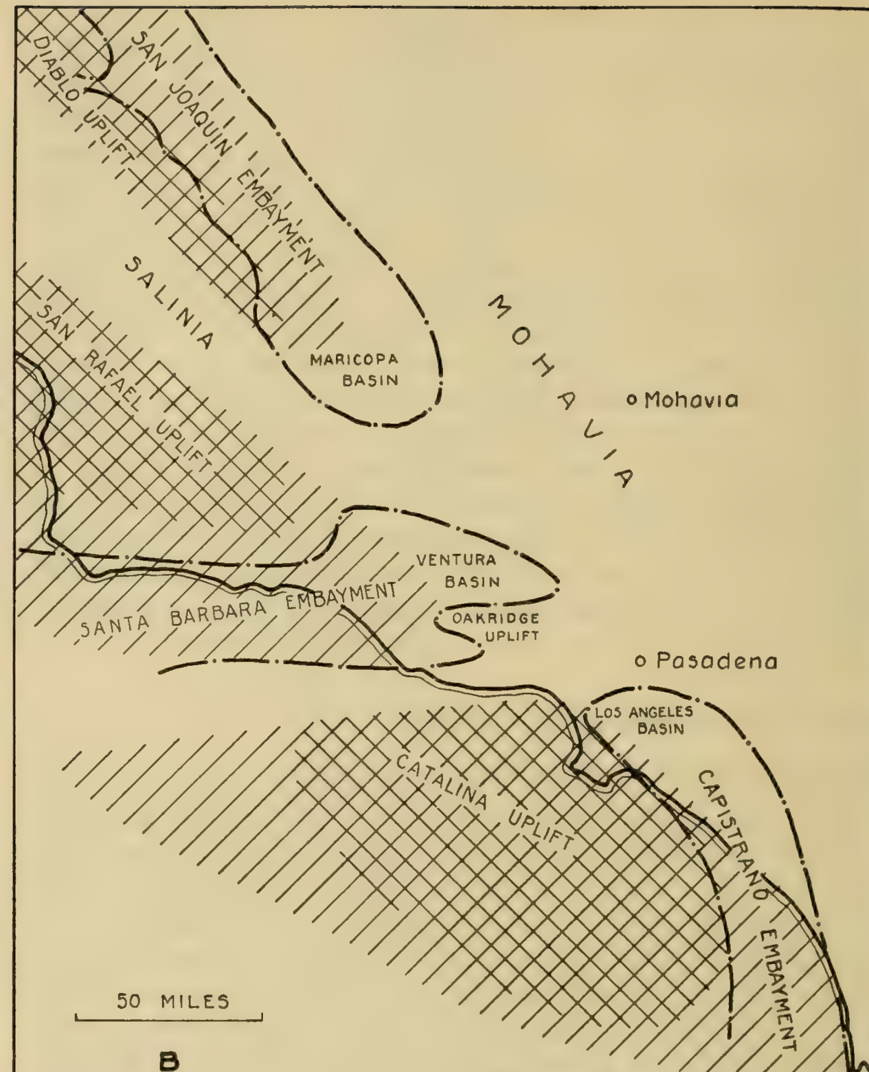


Fig. 29.3. A, map of southern California showing the distribution of Franciscan outcrops (black) and granite of the Nevadan orogeny (hachured). After Reed and Hollister, 1936. The granitic area marked *Salina* rose and became a landmass in Upper Cretaceous time, according to Taliaferro, 1943. Compare with Plate 16.

B, map of southern California showing the Tertiary provinces and their relations to the basement rock. See opposite map. After Reed and Hollister, 1936.

that the Coast Ranges of southern California have probably shifted some 300 miles northwestward to their present position, and therefore the paleotectonic maps of Fig. 29.3 are probably not correct. They show, however, the principal tectonic elements, and are reproduced because they help in understanding the make-up of the region.

The evolution of the central Coast Ranges in Early Tertiary time is idealized in Fig. 29.4, and the deposits under that part of the San Joaquin embayment that were not deformed appreciably and later became the San Joaquin Valley are shown in Fig. 29.5.

Early Tertiary Phase

In Chapter 17 it was pointed out that the Santa Lucian orogeny was the last disturbance in the Cretaceous, and that following it, the widespread, thick Asunción group was deposited in Senonian, Maestrichtian, and Danian time. The deposition of the Asunción was brought to a close in the Central Coast Ranges by uplift, tilting, and probably folding; but so little of the Paleocene is preserved that its original extent and thickness and the degree and extent of the post-Cretaceous disturbance are not known. However, it is believed that the disturbance was not as severe as previous ones, because the uppermost Cretaceous and Paleocene, where observed, are only slightly discordant, and very little change in the character of the sediments is noted (Taliaferro, 1943b).

Taliaferro (1943a) suggests that the Paleocene represents a final stage in the history of the late Mesozoic geosyncline, the California trough of this book, in which the Franciscan, Knoxville, Shasta, and Chico sediments were deposited. See Fig. 29.6. The part of the trough in the site of the present central Coast Ranges and along the western border of the present San Joaquin Valley received sediments throughout the late Upper Cretaceous, and then weak uplift, folding, and erosion occurred to the west, while the central part of the trough was little affected. Probably general uplift occurred, and the seas retreated; but the uplift appears to have been quickly succeeded by downsinking, and the Paleocene sea flooded at least parts of the Cretaceous deposits. This was the last time that deposition took place over rather large areas of the trough. The changes that had taken place previously were of lesser magnitude than

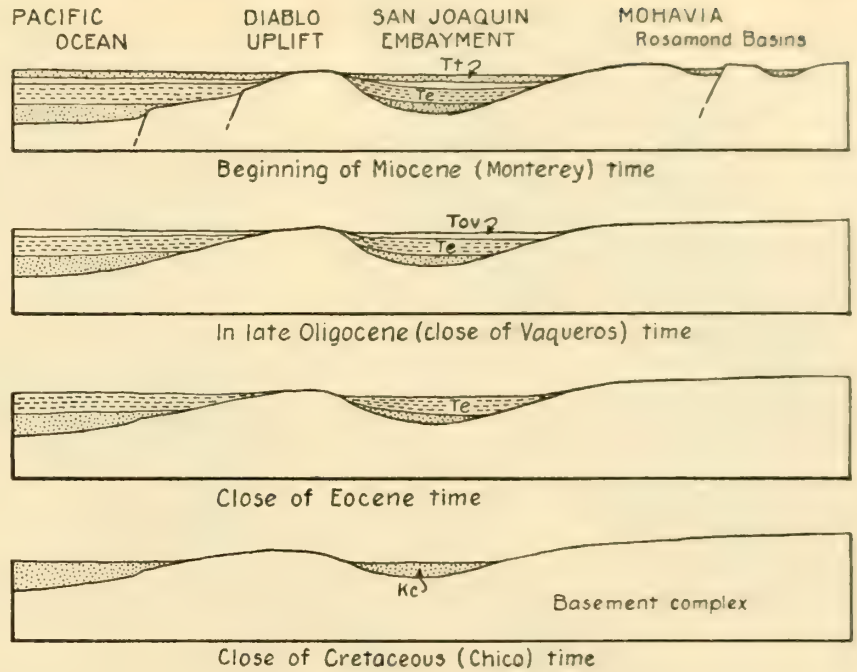


Fig. 29.4. Evolution of Coast Ranges and Great Valley in Early Tertiary time. Kc, Chico formation; Te, Eocene formations; Tov, Oligocene and Vaqueros formation; Tt, Temblor formatoin. After Reed, 1933. Section E–E′, Fig. 29.1.

those that took place *after* the Paleocene. The available evidence indicates that the final fragmentation of the California trough took place in the Eocene. Great thicknesses of Tertiary sediments accumulated, but they formed in comparatively narrow basins, some of which were at a marked angle to the more extensive and enduring late Mesozoic trough.

Eocene and Oligocene strata have limited distribution in the central Coast Ranges, and their nomenclature and correlation have been the objects of considerable discussion. Typical Tertiary formations are listed in the chart on page 458. Although thick lower Eocene sections occur, they are in small, isolated localities; and more of the California Coast Ranges were emergent than at any time during the Cretaceous and Juras-

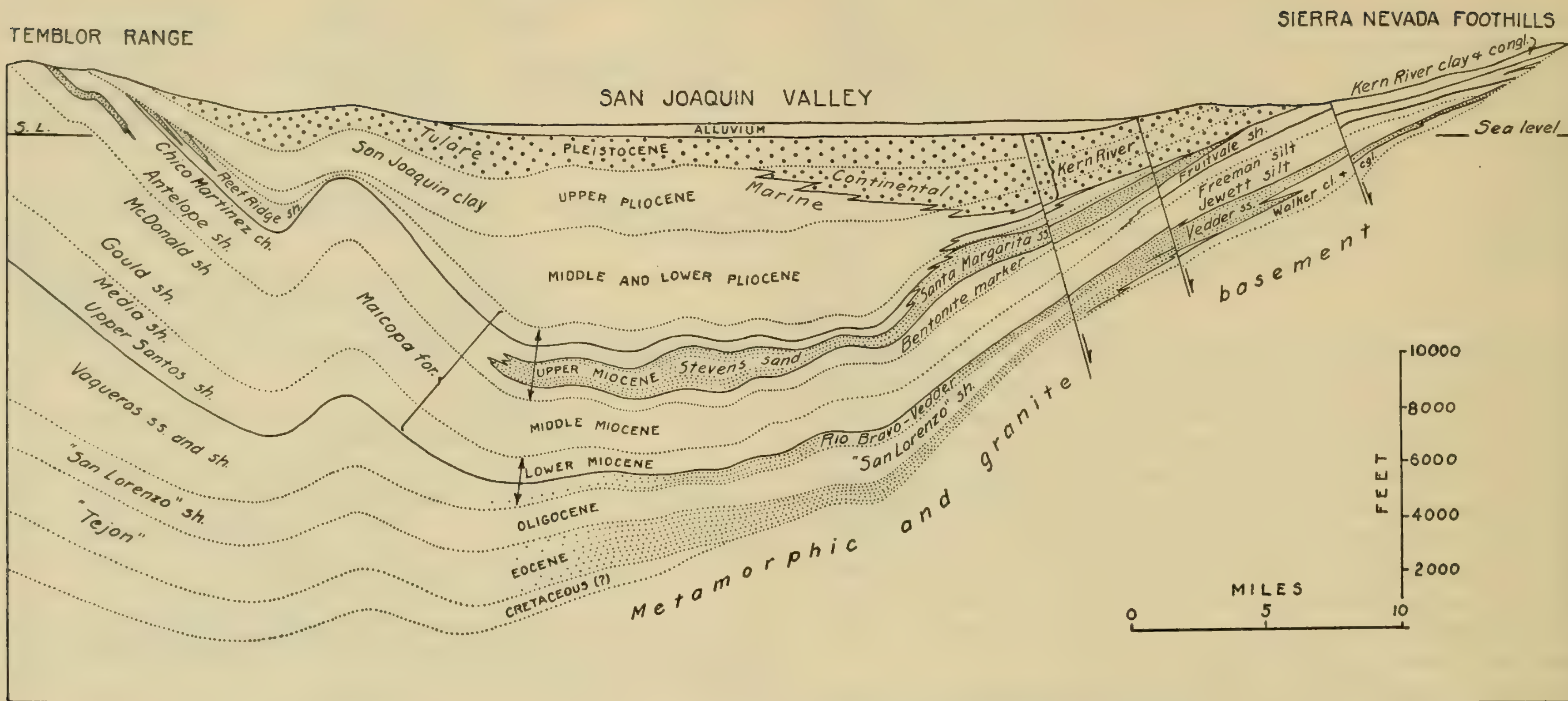


Fig. 29.5. Generalized section across the southern San Joaquin Valley. After Hill and Eckis, 1943. Section K–K′, Fig. 29.1.

sic. The Santa Lucia Range, most of the Santa Cruz Mountains, and much of the Diablo Range stood above sea level, as did also the central Sierra Nevada.

The middle Eocene sea appears to have had a much wider extent and to have flooded much of the San Joaquin embayment. The middle Eocene formations are recognized by Taliaferro (1943b) to be the Capay, Domengine, and Ione. They are sandstones, shales, clays, limestones, and coal beds; and they are unusually fine grained except at the margin of the border lands. The Ione is clearly of an eastern source, but the Middle Eocene along the Diablo Range contains detritus from the Franciscan and Cretaceous strata of the ancestral Coast Ranges as well as the crystalline rocks of the ancestral Sierra Nevada. The Middle Eocene covered the east flank and northern end of the Diablo Range, probably a part of the Santa Cruz Range, and northeastern part of the Santa Lucia Range. Minor volcanic activity can be recognized by rhyolitic and andesitic debris in the Ione of the Great Valley, supposedly of an eastern source in the Sierras, and by bentonite in the Domengine of the Coast Ranges, supposedly of a western source (Taliaferro, 1943b). See accompanying chart.

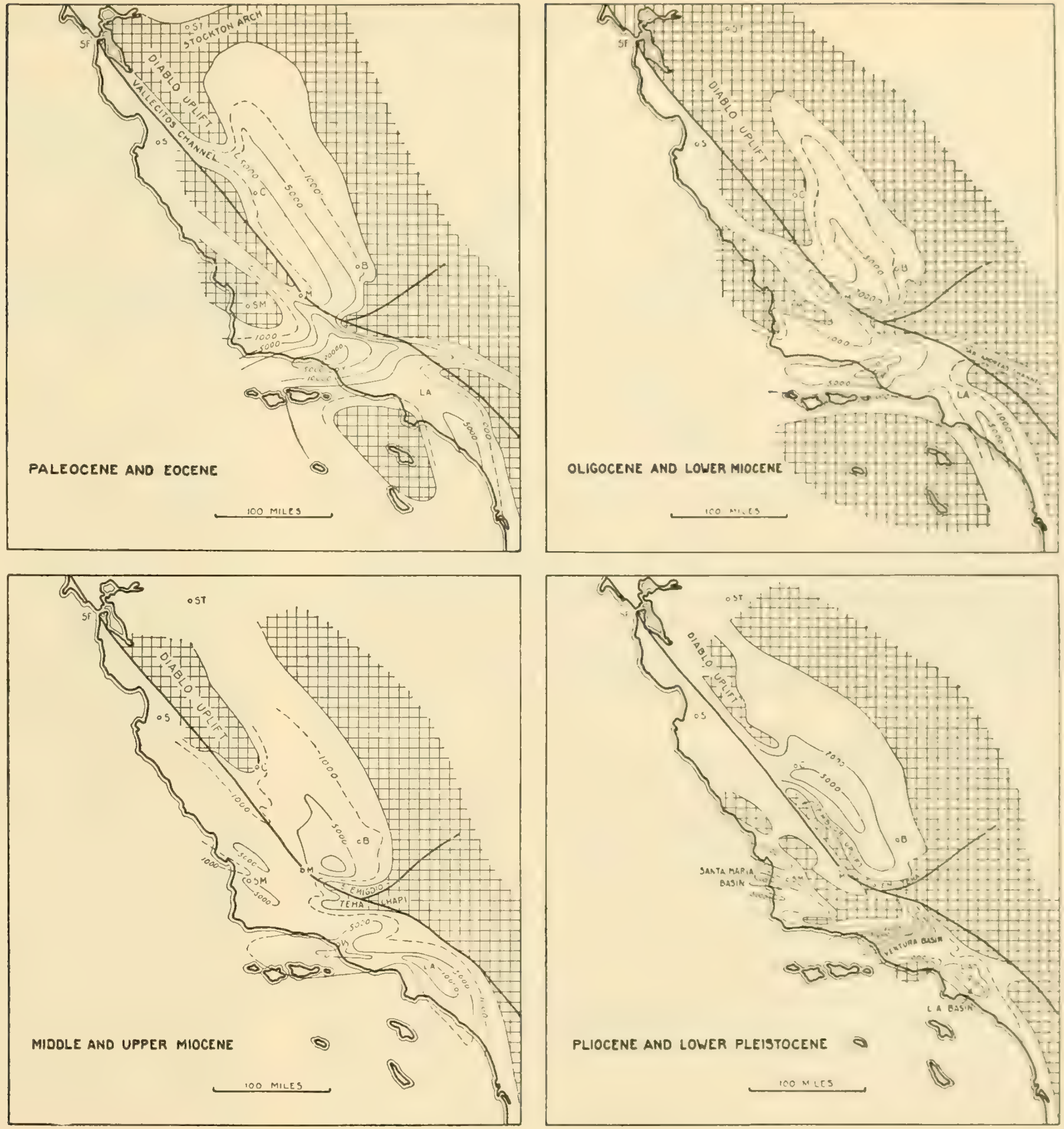


Fig. 29.6. Maps of southern California showing the basins of deposition and the land areas (cross-ruled) during the Tertiary. After Reed (1933) and Hoots *et al.* (1954). Compare these maps with those of Fig. 29.3. SF, San Francisco; ST, Stockton; S. Salinas; C, Coalinga; SM, Santa Maria; M, Maricopa; B, Bakersfield; V, Ventura; LA, Los Angeles. If the Coast Ranges southwest of the San Andreas fault have moved about 200 miles to the northwest since the beginning of Tertiary time, then progressive adjustments in its relation to the basins and lands on the northwest must be visualized.

Typical Formations in California	Age Assignments: Current Usage	Age Assignments: Grouping by R. D. Reed	
Upper San Pedro	Upper Pleistocene	Pleistocene	
Lower San Pedro, Saugus, Tulare	Lower Pleistocene	Upper Neogene	Neogene
Etchegoin, Pico, Repetto	Pliocene		Neogene
Santa Margarita, Monterey, Modelo, Topango, Temblor	Upper and Middle Miocene	Lower Neogene	Neogene
Vaqueros, Temblor, Pleito, San Lorenzo, San Ramon	Lower Miocene and Oligocene	Upper Paleogene	Paleogene
Kreyenhagen, Tejon, Capay, Domengine, Meganos, Martinez, Ione, Poway	Eocene and Paleocene	Lower Paleogene	Paleogene

The upper Eocene (Tejon, Markley, Kreyenhagen, Gaviota, and Wheatland) has a more limited distribution than the Middle Eocene. Very slight folding and faulting may have intervened, but no mountains were built, and the same seaways as before persisted, though somewhat restricted. The Kreyenhagen has some bentonite and vitric tuff beds, and the Wheatland has some andesitic debris, both indicating continued small-scale volcanic activity.

The Oligocene strata have even a more restricted distribution than the upper Eocene, but occupy the same basins. They generally rest unconformably on Eocene sediments and, in turn, are generally unconformably overlain by the Miocene. The sediments regarded as Oligocene at present are those of the San Lorenzo group. Volcanism occurred during the Oligocene in the Mount Diablo and San Francisco Bay regions, where more than 100 feet of rhyolite tuff occurs in the Kirker formation.

The disconformities and slight angular unconformities that are known in the Eocene and Oligocene might indicate comparative quiet in strong contrast to the preceding and succeeding periods. This seeming lack of important diastrophism, however, may be more apparent than real because of lack of evidence. The Upper Jurassic and Cretaceous unconformities show that the various crustal movements were strongest in the western coastal region, the volcanic archipelago, and died out eastward. The same may be true of the Eocene and Oligocene (Taliaferro, 1943b).

The structures formed probably represent the general effect of several episodes of movement. Although both folding and faulting occurred, normal faulting in the Diablo Range of great magnitude predominated. It was during the Early Tertiary phase that the uplift and westward tilting of the Gabilan Mesa (Diablo uplift) occurred, approximately along a line corresponding to the present position of the San Andreas fault. This northeastern boundary fault may be thought of as ancestral to the San Andreas fault in the central Coast Ranges, where the two coincide. The southwestern side of the uplift is irregular, with several smaller faults. See Fig. 29.4.

Late Miocene Phase

Over most of the central Coast Ranges, the Miocene began with gentle sinking, and basins of the early Tertiary were first uniformly flooded and then overlapped. Early in middle Miocene, the uniform and gentle sinking gave way to sharper downwarping, and great thicknesses of sediments accumulated locally. It is believed that the movement was caused by compression and that the interbasin areas rose at the same time as the basins sank. The heterogeneous pre-Tertiary basement is believed to have precluded uniform folding throughout the Coast Ranges. An important and rather long-enduring trough developed along the western downtilted side of the Gabilan Mesa, west of the Santa Lucia Range. The trough east of the range continued to sink and expand both southward and northward, until a connection was made with the sea in the site of the present Monterey Bay.

The crest of the Coalinga anticline, now composed of Franciscan, stood above sea level throughout the Miocene.

The effect of movements during the later upper Miocene cannot be clearly evaluated in all places, because erosion incident to later severe deformation has removed much of the evidence. This is especially true in the Santa Lucia Range. However, in the northern part of the Castle Mountain Range, the nature of upper Miocene deformation is well shown. Figure 29.7 has been prepared to illustrate the structural evolution. Santa

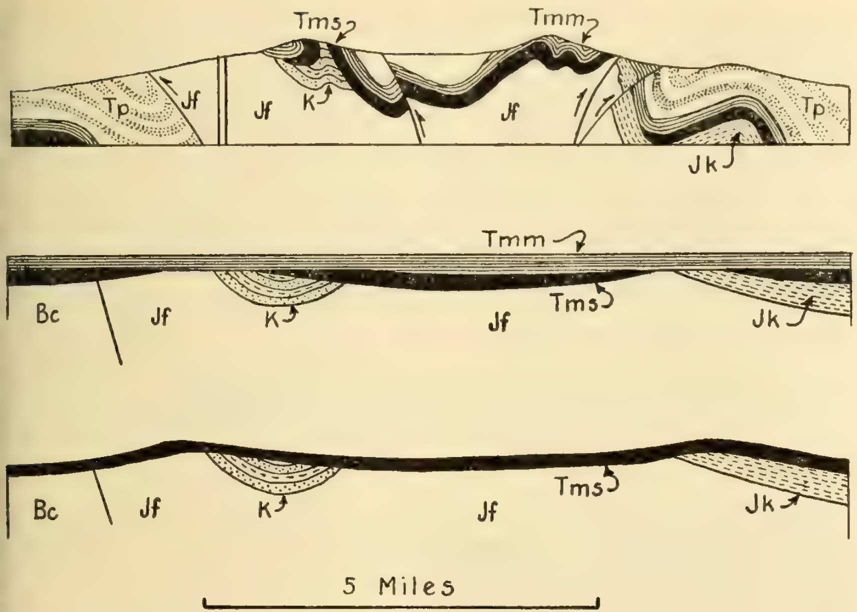


Fig. 29.7. Evolution of the Castle Mountain Range. Ideal sections showing late Upper Miocene folding (lowest section), erosion and deposition of McLure shale (middle section, latest Upper Miocene), and thrusting and folding in late Pliocene (upper section). The marginal thrusts of the Castle Mountain Range developed in the sites of the Upper Miocene anticlines. More thrusting occurred in mid-Pleistocene which is not represented. After Reed and Hollister, 1936. Section H–H′, Fig. 29.1. Tms, Santa Margarita sandstone; Tmm, MsLure shale; Tp, Purisma fm.

Margarita sands 100 to 300 feet thick of late Miocene age were deposited over a fairly even-floored basement complex consisting chiefly of Franciscan, but with remnants of Knoxville, Shasta, and Upper Cretaceous sediments. After, or perhaps even during the deposition of these sands, gentle anticlinal folding occurred along two subparallel lines 6 to 8 miles apart, which correspond approximately to the present margins of the range. The maximum observed dip of the flanks is 11 degrees. The two anticlinal ridges were planed off, perhaps as rapidly as they rose, and the McLure shale was then deposited over the region. Where it crosses the two anticlines, it lies unconformably on the Santa Margarita sands and on the Franciscan. Elsewhere, the Santa Margarita and McLure are conformable, and in places they appear to grade into one another. In the later Pliocene and Pleistocene deformation, thrusts developed approximately in the sites of the anticlines.

Late Pliocene and Mid-Pleistocene Phases

The thick accumulation of Miocene sediments was accentuated, in general, by further deposition in the same troughs in early and middle Pliocene time. The gentle compressive movements, which started in the Miocene and then relaxed for a while, surged to a peak in the late Pliocene and again to another peak in the mid-Pleistocene. The last surge is probably still climactic.

The folds and thrust faults that are the conspicuous features of cross sections and field observation are largely the result of these two movements. Cross sections D–D′, F–F′, G–G′, and I–I′ of Fig. 29.8 are especially illustrative of the compressional deformation to which the rocks of the Diablo uplift and the San Joaquin embayment in the central Coast Ranges were subjected.

Opinions differ as to the relative importance of the two phases. In some places, only one has been recognized. According to Taliaferro (1943b), the geologists in general who have worked in the western part of the Coast Ranges have emphasized the importance of the late Pliocene disturbance there, and those who have worked chiefly in the eastern part have stressed the mid-Pleistocene compression.

The regions underlain at comparatively shallow depths by crystalline rocks, or those where the crystalline rocks were exposed, yielded by faulting; and those underlain by thick sections of strata (8000–20,000 feet) yielded by folding and thrusting (Taliaferro, 1943b), except for the eastern part of the San Joaquin embayment which was left little deformed and is now the Great Valley. The ranges are generally bordered by thrusts, but the individual thrusts can be traced only 20 to 25 miles. As a thrust dies out, its place is commonly taken by one or more *en echelon* faults. The thrusting is both westward and eastward, with some structural units (ranges) being bordered by complementary inward-dipping faults. The thrusts marginal to the structural units generally have shallower dips than those within. The structure of the Central Coast Ranges as interpreted in the cross sections is rather similar to that of the Montana and Alberta

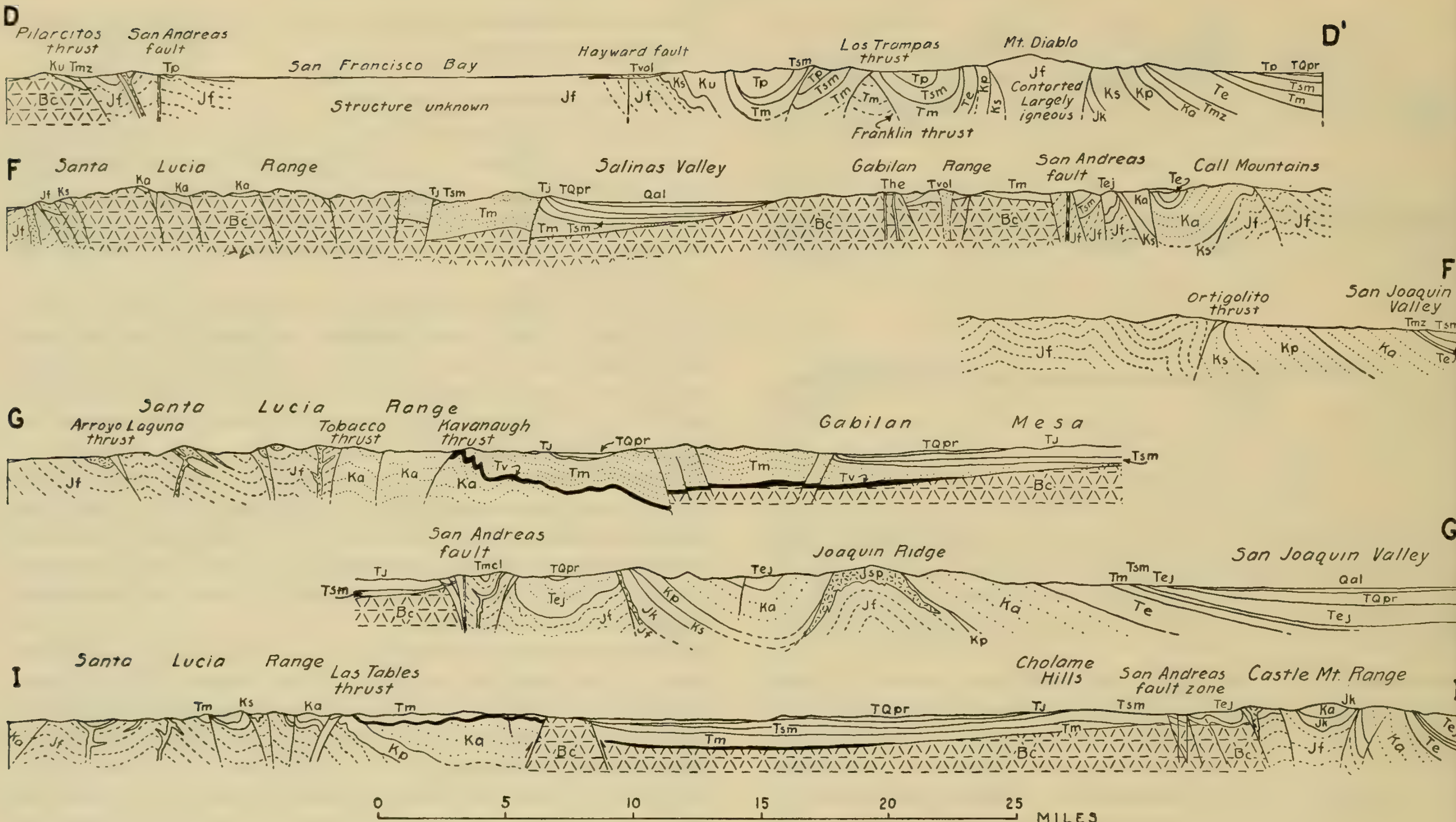


Fig. 29.8. Cross sections of the central Coast Ranges after Taliaferro, 1942. Refer to index map, Fig. 29.1. Bc, granite, gneiss, schist, and marble; Jf, Franciscan sediments and volcanics; Jk, noxville shales and sandstones (Jf and Jk are Upper Jurassic); Ks, Shasta group, Lower Cretaceous; Kp, Pacheco group; Ka, Asuncion group; Ku, undifferentiated (Kp, Ka, Ku, Upper Cretaceous); Tmz, Martinez, Paleocene; Te, Eocene undifferentiated; Tv, Vaqueros, Lower Miocene; Tm, Salinas shale, Temblor, etc., Middle Miocene; Tmcl, McLure shale, Upper Miocene; Tsm, Santa Margarita, San Pablo, etc., Upper Miocene; Tvol, volcanics, sills, dikes, Miocene; Tej, Etchegoin, Tj, Jacalitos, Tp, Purisma, Pliocene; TQpr, Paso Robles, Santa Clara, San Benito, Tulare, etc., Plio-Pleistocene.

Rockies. The Coast Ranges have a more heterogeneous basement, which has served to localize the thrusts; the strata in them are generally less indurated; the scale is somewhat smaller; and the movement along the thrust surfaces is generally less.

In addition to the thrusts, there are transverse faults, some of which cut almost completely across a range. They relate to the uplift of the Santa Lucia Range, because in its southern part, each transverse fault is downthrown on the south, and the range becomes progressively lower in elevation in that direction (Taliaferro, 1943b).

There was little volcanism in the central Coast Ranges during the Pleistocene as compared with the extensive and important volcanism in the Sierra Nevada Range and in the Cascades. Olivine basalt flows and agglomerates occur in the Santa Lucia and Diablo ranges and along the east side of Santa Clara Valley.

The Tertiary structural history was much like that of the Late Jurassic and Cretaceous in the following respects. The Orogeny was generally severest westward, because the unconformities are more angular and bring rocks of greater age differences together the farther west from the Great Valley they are observed; and volcanism continued, with tuffs and flows a characteristic part of middle Eocene, upper Eocene, lower Miocene, middle and upper Miocene, lower and middle Pliocene, and Pleistocene formations.

The mid-Pleistocene orogeny occurred farther inland (eastward) than the late Pliocene orogeny and is a contrary note to the generalization of increasing intensity westward. However, the two disturbances are closely connected in time and may be part of a general wave of deformation originating in the west and progressing eastward.

The mid-Pleistocene disturbance is associated with the final disappearance of the Tertiary troughs of deposition and the foundering of considerable segments of the Coast Ranges into the Pacific. It is evident from inspection of the tectonic and geologic maps that the sea has transgressed part of the Coast Range orogenic belt; the structures are discontinuous at the present shore line. Also, the reconstructed Tertiary uplifts and troughs head out to sea, as if only half exposed in the Coast Ranges. Recent detailed mapping of the ocean floor off California has revealed a topography much like that in the Coast Ranges, and it can best be explained as the surficial expression of the long-evolving volcanic archipelago of Paleozoic, Mesozoic, and Cenozoic time, with particular respect to the late Pliocene and Pleistocene deformations. The interpretation of the topography of the sea floor will be taken up later in a separate chapter.

The San Andreas fault, that stretches through the three divisions of the Coast Ranges of California will be considered later.

Erosion Following Main Orogeny

Following the late Pliocene and mid-Pleistocene orogeny, which resulted in rapid uplift and oversteepening of the mountain fronts, vigorous erosion reduced the escarpments and ranges until now there is little physiographic evidence left of individual faults, although some of them were of several thousand feet displacement. Conspicuous features of the rapid erosion are the landslides from the oversteepened mountain fronts. Some were gigantic in size and took place coincident with the thrusting and uplift of the ranges; others have occurred since. In places, there is a definite sequence of slides observable, detected by different amounts of dissection. They obscure the true structure of the mountain front in many places.

Late Pleistocene and Recent Gentle Folding

In the Los Angeles, Ventura, and San Joaquin basins, gentle folds have developed so recently that they have been little modified by erosion, and precise elevation surveys show that movement is still going on vigorously. The subject will be taken up at greater length under the next major heading, "Southern Coast or Transverse Ranges of California."

Terraces

Terraces are numerous and well developed along the shore and in interior valleys of the Coast Ranges. The marine terraces are found at elevations up to 1500 feet, and attest the rise of the Coast Ranges in very recent times. They are cut on the beveled edges of the folded Plio-Pleistocene sediments, and therefore are very young. Individual terraces are difficult if not impossible to follow from one region to another, and

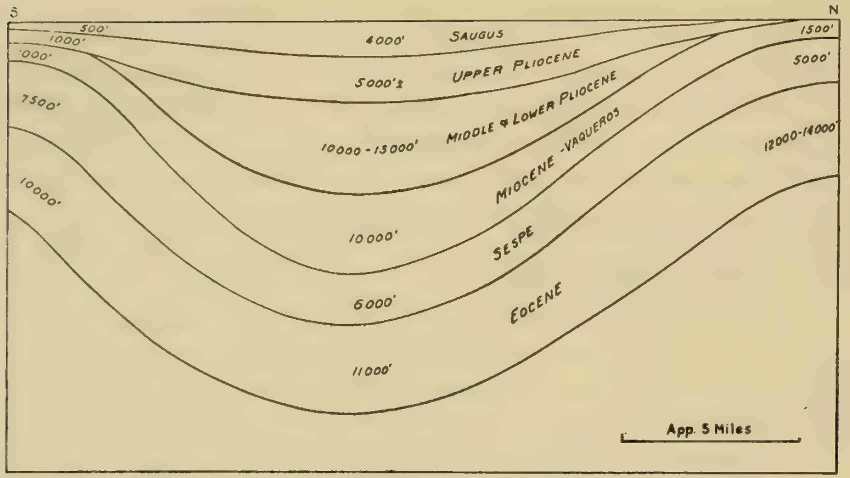


Fig. 29.9. Ventura basin showing conditions before Middle Pleistocene folding. Section P–P′, Fig. 29.1.

there is little definite correspondence of the various terrace levels over wide areas. Over limited areas, there may be very definite intervals between terrace; a few miles away, the terraces may be equally well developed; but the intervals between terraces in the two areas differ. Furthermore, the marine terraces along the coast cannot be correlated definitely with the terraces of the interior valleys, but there is strong evidence that the coastal area has very recently been uplifted more than the interior (Taliaferro, 1943b).

The San Francisco Bay area was probably depressed rather than uplifted, but it is not possible to say that the entire lowland and bay was depressed subsequent to the folding and thrusting because it may have been left that way as orogeny progressed (Taliaferro, 1943b).

The terraces have been cited as evidence of widespread epeirogeny, but Taliaferro thinks they may be due to gentle folding or upbowing of the ranges.

A few but indisputable examples of tilted beaches are known, but the structural meaning is yet obscure.

SOUTHERN COAST OR TRANSVERSE RANGES OF CALIFORNIA

Principal Structural Features

The Southern Coast Ranges trend in an east-west direction which is transverse to that of the Central and Peninsular Ranges. See Fig. 29.2. The relief features as well as the faults and folds are generally so oriented.

The formations and structure of the southern part of the San Joaquin basin are shown in cross section in Fig. 29.5. A cross section of the Ventura basin, restored to the time preceding the major deformation, is presented in Fig. 29.9. The cross sections L–L′, M–M′, and N–N′, Fig. 29.10, and O–O′, Fig. 29.11, are representative of the present structure and major groups of beds in various parts of the southern Coast Ranges.

Early Tertiary Phase

Paleocene, Eocene, Oligocene, and early Miocene times were generally characterized by subsidence of the basins previously mentioned, but at times during these epochs slight surges of crustal unrest are attested by conglomerates and local small-angle unconformities. During the Eocene, the greatest subsidence occurred, and it centered in the Ventura basin. The Paleocene beds are generally coarse, variable in lithology, and of restricted distribution. In most places, the contact with Cretaceous beds is difficult to locate, and the two systems seem conformable. Aside from the coarser aspect, the Paleocene beds are not much different from the Cretaceous. In one locality, an angular unconformity of 30 degrees has been noted (Reed, 1933), and it has been taken to mean gentle folding in places at the beginning of the Tertiary.

The Eocene sediments were generally finer, and consisted of arkosic sandstones and silty and sandy shales. They accumulated to a depth of 11,900 feet in the Ventura basin. Perhaps the total thickness there of Paleocene and Eocene beds, the Martinez and Tejon formations, was 20,000 feet. See thickness contours of Fig. 29.6. The Eocene deposits spread over much larger areas than the Paleocene, but the subsidence followed the earlier troughs or defined them better. Toward the end of the Eocene or during the Oligocene, the areas of deposition remained large, but the facies represented became highly varied. They included the Poway

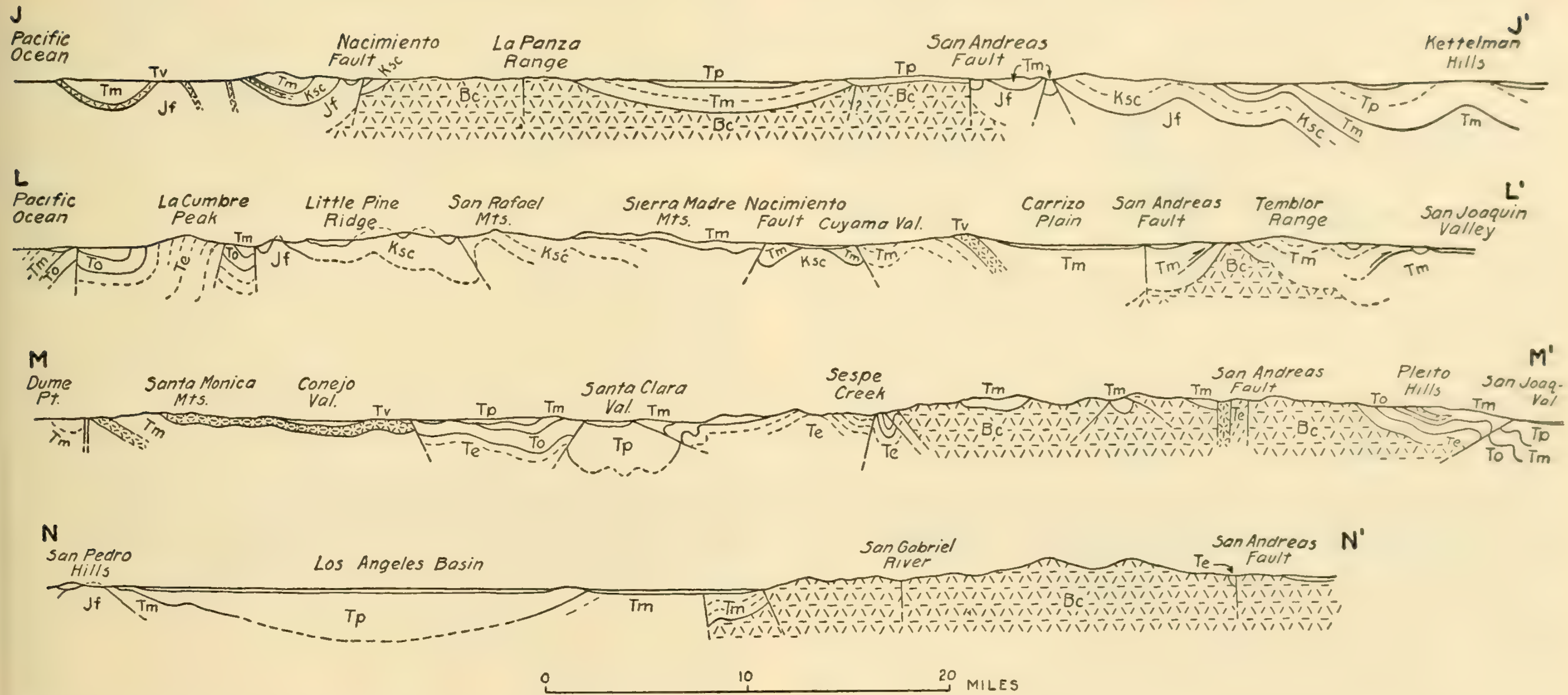


Fig. 29.10. Generalized cross sections of the Coast Ranges of southern California, after Reed and Hollister, 1936. Bc, granite and metamorphic rock basement; Jf, Franciscan; Ksc, Cretaceous strata; Te, Eocene strata; To, Oligocene strata; Tm, Miocene strata; Tp, Pliocene strata; Tv, volcanics. Sections J–J', L–L', M–M', and N–N' of Fig. 29.1.

conglomerate, lower Sespe continental red sandstone, Coldwater and Tejon marine sandstone and sandy shale, and the Kreyenhagen siliceous shale.

Oligocene and early Miocene (Fig. 29.6) time saw a great increase in size of the land areas, considerable parts of which received thick deposits of red and green shales, the nonmarine part of the Sespe formation. Later on in Vaqueros time, the sea invaded much of the Sespe lowland. In the San Joaquin embayment, the Kreyenhagen shale was deposited, and it graded into sandstone southward. Still farther southwest, in the Santa Barbara embayment, the Sespe red beds accumulated. Reed (1933) believes a basin had become semi-inclosed and was gradually filled with silts and oozes of high organic content that later evolved the oil in the Coalinga district.

Although the land areas increased in size and the seaways decreased, the Santa Barbara trough continued strongly negative, and Miocene and Pliocene sediments accumulated 25,000 to 30,000 feet thick in the deepest

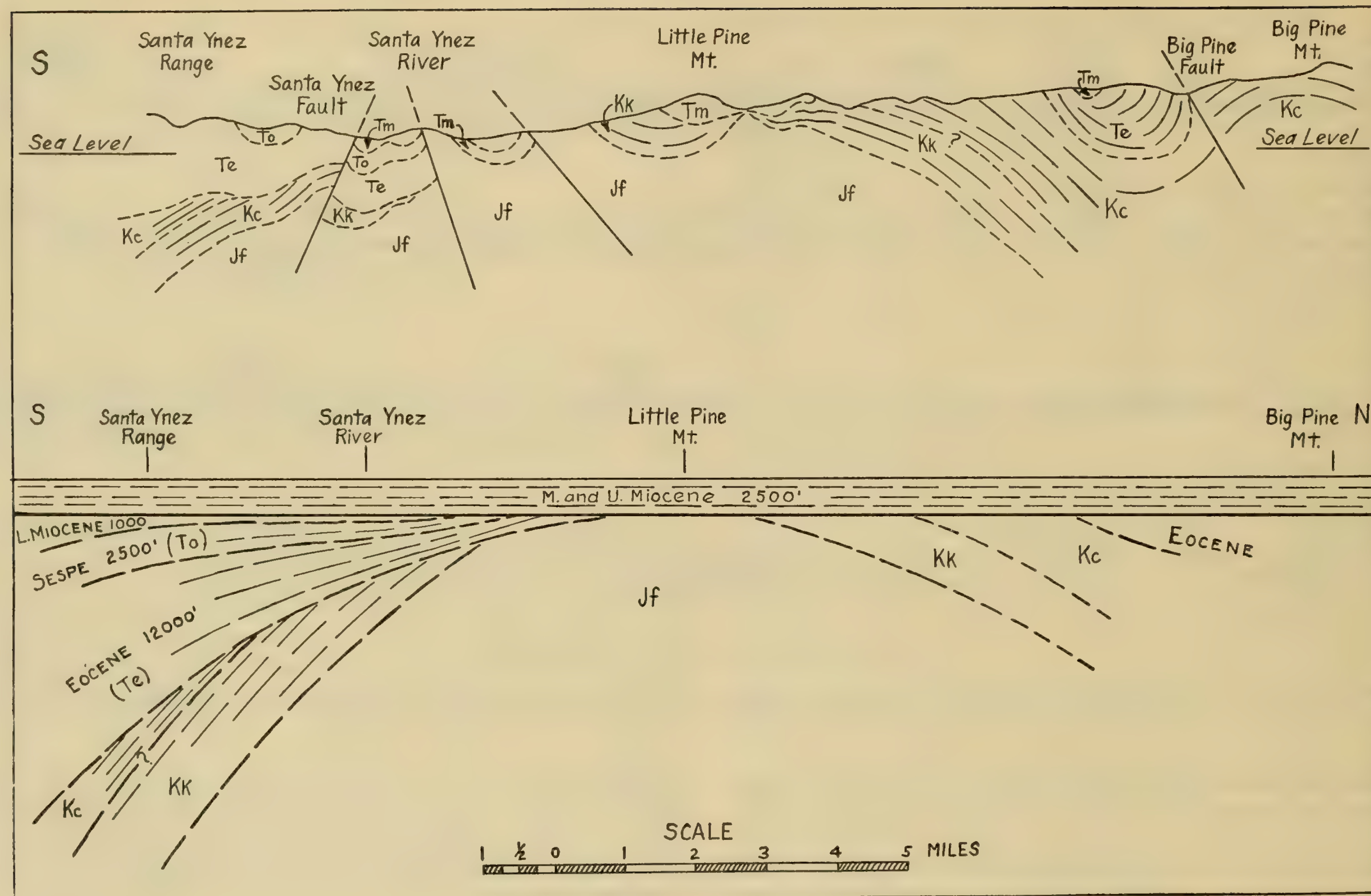


Fig. 29.11. Present and Upper Miocene structure in Santa Ynez–Santa Barbara district. Section O–O', Fig. 29.1. Jf, Franciscan; Kk, Knoxville; Kc, Chico, Te, Eocene formations; To, Oligocene formations; Tm, Miocene. Reproduced from Reed and Hollister, 1936.

part. The Oakridge uplift started to evolve when, at the end of Sespe time, local subsidence ceased to be so rapid. The lower Miocene strata include the Vaqueros sandstone and the Rincon (Temblor) clay shale. The middle Miocene has a basal limestone. Both middle and upper Miocene contain a predominance of siliceous organic shale, including diatomite, chert, and various other siliceous varieties.

Middle and Late Miocene Phase

The late Miocene phase (Fig. 29.6) is not well known in the southern Coast Ranges, and the evidence that is available suggests only local, gentle folding and volcanism. Near Santa Barbara, a coarse breccia of middle late Miocene age occurs at the plunging ends of cross folds in the east-west trending structures. Reed believes this breccia was formed during the cross folding. In the cross section O–O′, Fig. 29.11, the folding seems to be mostly pre-middle Miocene, and the anticline grew by several movements from Eocene to Miocene. It is probable that the Santa Barbara district of the Santa Barbara embayment in middle Miocene time was one of southward regional dips, fluted by a few low folds of northerly trend. The most important of the cross folds was along the axis of the San Rafael uplift and extended to Ventura (Reed and Hollister, 1936). East and west of Los Angeles, in the Los Angeles basin, gentle folding occurred in late Miocene time. In spite of the folding and the change to heterogeneous facies in the late Miocene from homogeneous facies in the middle Miocene, the boundaries of the two basins were much the same (Reed and Hollister, 1936).

Volcanic rocks form an important constituent of the middle and upper Miocene along the axis of the San Rafael uplift but are not conspicuous elsewhere.

Pleistocene Phase

Pliocene and lower Pleistocene deposits of considerable thickness occur only in restricted parts of the Miocene basins. See Fig. 29.6. The three thickest deposits of Pliocene beds in southern California are found in the Maricopa, Ventura, and Los Angeles basins. The thicknesses are very great, possibly 10,000 to 20,000 feet in the first, 18,000 feet in the second, and 10,000 in the third. The foraminifera in the lower Pliocene beds of the Ventura and Los Angeles basins suggest that the sea was one to two miles deep at the time of deposition.

Lower Pleistocene was deposited in all the Pliocene basins, but only in the western depressions did marine beds accumulate. Eastward, the beds are continental and are 1000 to 5000 feet thick.

The late Pliocene orogenic phase of the central Coast Ranges as described by Taliaferro is not a "notable disturbance" in the southern Coast Ranges, according to Reed and Hollister (1936). A disconformity is present in the Repetto Hills, the Ventura basin, and the San Joaquin Valley between the lower and upper Pico, but the break has not been observed as an angular unconformity anywhere.

Along the seaward margin of the Los Angeles basin, there is a pronounced angular unconformity between lower San Pedro beds of early and middle Pleistocene age and upper San Pedro beds of late Pleistocene age (Reed and Hollister, 1936). The evident folds and thrusts of southern California can best be explained, according to Reed and Hollister, as having formed approximately at this time. Examine cross sections L–L′, M–M′, N–N′, Fig. 29.10, and O–O′, Fig. 29.11.

In the Ventura basin, about 5000 feet of lower Pleistocene beds have been turned up so as to have dips of 30 to 90 degrees beveled by erosion and covered by about 300 feet of upper Pleistocene fanglomerates. The fossil *Equus*, cf. *occidentalis*, occurs both above and below the angular conformity, apparently without change, and indicates that the structure was formed during a very short period of time (Bailey, 1943).

The structural history of the Kettleman Hills anticline is instructive. In it, the Tulare formation, which is lower Pleistocene and not older than latest Miocene, is folded apparently as strongly as the underlying formations. The anticline, therefore, was formed almost entirely in post-Tulare time. After its rise, it was eroded until several thousand feet of rock were removed from its axial part. Toward the south end, it was reduced to a plain which then became buried in alluvium. After this, the alluvium was arched into a new, though gentle, fold.

In the Los Angeles basin, a number of unconformities within the upper Miocene and the Pliocene section indicate a succession of uplifts along the major structural trends during these times (Wissler, 1941).

The present condition of the crust in southern California is one of decided instability. Folding, thrusting, and high-angle faulting have not only manifested themselves in earthquakes and buckled pipelines, cables, and pavements; through precise surveys, the amount and rate of the movements have been measured in places. Gilluly (1949) reviews these movements and concludes that the present is a time of typical orogeny. The seismicity of the western Cordillera will be considered in Chapter 31.

It is evident that the division of the structural history of southern California into phases is not altogether a satisfactory treatment, because the deformation was prolonged and shifting in time and place. Basin subsidence, sediment accumulation, the tilting and erosion of marginal beds, and the rise and truncation of anticlines in nearby and related areas all went on together. Very little time is represented in some of the angular unconformities, hardly enough for a change to occur in the faunas, yet the angular unconformities have caught and fixed the rise of landmasses in process of movement in the same manner almost as a photograph stops an object in motion. Perhaps the unconformities should not be considered rigidly as indicators of separate widespread impulses. In the analyses of Paleozoic and Mesozoic orogenies in the great system of western Cordilleran troughs, the theory seems repeatedly substantiated that deformation was almost continuous in an oceanward volcanic archipelago, and that from time to time the compressive movements spread into the flanking trough and deformed the sediments in it to variable intensities and distances. These deformational waves off the main belt of constant unrest probably constitute our orogenic impulses or phases in the Coast Ranges.

NORTHERN COAST RANGES OF CALIFORNIA

General Features

The Northern Coast Ranges, as generally defined, extend from San Francisco Bay to Trinidad Head and perhaps beyond. They are bounded on the east by the Sacramento Valley and on the north by the Klamath Mountains. They are composed mostly of Franciscan-Knoxville strata, but other pre-Tertiary formations may be present; and in this respect, they contrast with the Central and Southern Coast Ranges, which in good part are made up of Tertiary deposits.

The southern end of the Northern Coast Ranges is not greatly different from the northern end of the Central Coast Ranges. In both the Tertiary is prominent, but northward it is limited to a few small basins and to the marginal areas. Most of the hills and valleys are probably underlain only by Mesozoic rocks. The complex structure of the Mesozoic and Cenozoic rocks, their poor outcrops in many places, and their slight economic importance as yet, have contributed to a lack of detailed geologic work except in a few areas.

Weaver (1949) has published on seven quadrangles north of San Francisco Bay, and reports that the hills there are arranged in three blocks, one west of the San Andreas fault, the Montara block; one east of it and west of the Tolay fault (a northwestward extension of the Haywards fault system), the Francisco–Marin block; and one east of the Tolay fault, the Berkeley Hills block. Refer to Figs. 29.1 and 19.2. Each block is tilted toward the northeast. The Franciscan group constitutes the surface exposures in most of the intermediate block, and the eastern block is made up of more than 30,000 feet of Jurassic to Quaternary marine and fresh-water sediments, together with about 1200 feet of Pliocene andesites, rhyolites, and tuffs.

> These sediments probably accumulated in structural troughs whose areas and physical environments changed greatly during the Cretaceous and Tertiary. The lower portion consists of clay shales and subordinate amounts of sandstone and conglomerate as much as 17,000 feet thick, containing a marine fauna of ammonites, pelecypods, and gastropods. These rocks include the Jurassic and Lower Cretaceous portions of the Knoxville formation and the Upper Cretaceous Chico. Several faunal zones may be distinguished in the Knoxville, but the formation in the mapped area cannot be subdivided on a lithologic basis. The chico formation consists of interbedded shales and sandstones about 7000 feet thick.
>
> The Paleocene is represented by the Martinez formation, and the Eocene in ascending order by the Capay shale and the Domengine and Markley sandstones. The formations of the Paleocene and Eocene series consist of marine

sediments ranging in thickness from 2000 to 5000 feet that were deposited in embayments far more restricted in area than the seas of the Upper Jurassic and Cretaceous time. The marine sedimentary formations of the Oligocene and lower part of the Miocene series occupy still more restricted areas than those of the Paleocene and Eocene, and near Carquinez Strait are more than 5000 feet thick. The upper Miocene sandstones of the San Pablo group are far more widely distributed and are nearly 2500 feet thick. They are characteristically coarse-grained and were deposited in moderately shallow water which locally was brackish or fresh. The Pliocene rocks crop out extensively in the north-central part of the area and consist largely of alternating flows of andesite, basalt, dacite, and rhyolite together with associated tuffs and agglomerates, whose total thickness is 100 to 1200 feet. In Santa Rosa and Petaluma quadrangles marine sandstones contain invertebrate fossils closely allied to those of the Merced formation in San Francisco. The beds in Marin and Sonoma counties are about 250 feet thick and rest unconformably upon the Franciscan group. Near Petaluma Valley they interfinger with tuffs (Weaver, 1949).

The Eel River embayment north of Cape Mendocino is the largest area of Tertiary sediments, and the beds there are said to be 7000 to 11,000 feet thick and of Pliocene age. Another deposit extends along the coast at Point Arena, where Miocene beds are several thousand feet thick. A third deposit is near Clear Water Lake, where 4000 feet of lower Eocene beds have been identified.

Early Pliocene Phase

In early Pliocene time before the Pliocene volcanics accumulated, the entire area east of the San Andreas fault was folded and faulted, and then deeply eroded. Particularly a great low-angle overthrust, the St. Johns Mountain thrust fault, was formed at this time.

Late Pliocene and Quaternary Phases

The Pliocene volcanics were laid down on the beveled surface of the older rocks, and later were moderately folded and broken by normal faults and locally overturned and broken by thrust faults.

Since the Pliocene beds in the Eel River embayment (Fig. 29.2) are folded and faulted, the northern part of the northern Coast Ranges was deformed in late Pliocene and Pleistocene time. This phase is similar to that in the San Francisco Bay area on the south. The main middle area is undoubtedly structurally complex, but it seems reasonable to conclude that it also was folded in late Pliocene and Pleistocene time, and perhaps during earlier phases.

Late Pleistocene and Recent Movements

As in the central Coast Ranges, there have been significant elevatory movements since the compressional deformation. The movements seem to be vertical and horizontal along faults, and also broader elevatory and depressional warpings.

Perhaps long before the compressional orogeny, the Klamath Mountains area projected westward as a peninsula, with the flanking areas below sea, especially on the north and west. A widespread erosion surface is believed to have developed over the Klamaths during this time (Fenneman, 1931). Then during the folding and thrusting on the north and west, it was only elevated, the Klamaths standing like a buttress to the deforming belts of Cretaceous and Tertiary strata. In relation to the trough sediments, the borders of the buttress were pushed westward up and over them.

Broad valleys were then cut in the high Klamath surface, according to Fenneman (1931), but not in a single uplift because the valley walls are terraced, and locally the floors of these broad valleys are themselves fairly widespread erosion surfaces. The highest peaks in the Klamaths rise several thousand feet above these broad valleys. In the Coast Ranges proper, there are remnants of erosion surfaces, but they have probably been jostled about in fault block movements. Their age, although most probably post-folding and post-thrusting, is not clearly demonstrable nor easy to compare with the Klamath peneplain and the broad valleys cut in it.

A great uplift affected the Klamaths and adjoining areas after the erosion of the broad valleys. Deep inner valleys 1000 to 2000 feet deep were cut and later glaciated. As the glaciation is generally recognized as Wisconsin, it would follow that the uplift and high-erosion surface are pre-Wisconsin in age. The uplift of the Klamaths may have been associated with the adjacent compressional orogeny, or it may have followed closely. At any rate, the uplift and dissection must have occurred in middle or post-middle Pleistocene. The narrow continental shelf was added to the

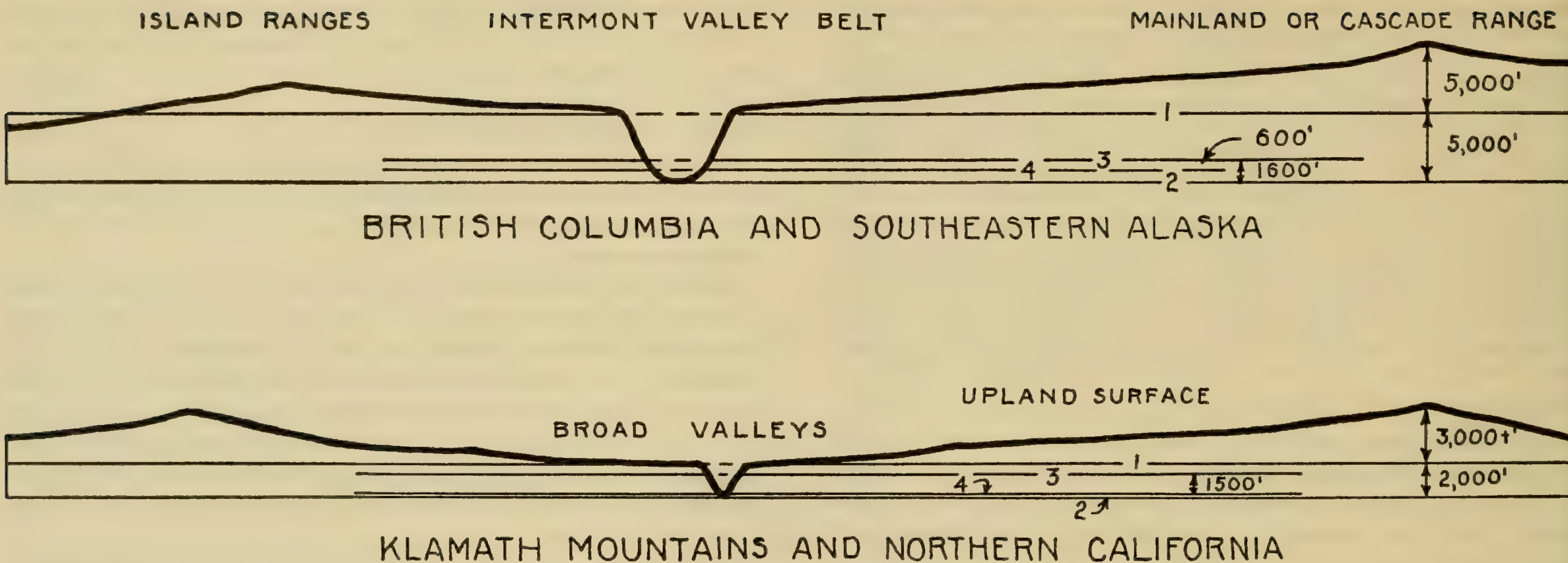


Fig. 29.12. Idealized diagrams to represent vertical movements of the crust in Pleistocene time along the Pacific coast. The upper diagram is schematic for the coastland of British Columbia and southeastern Alaska. It runs east-west, and the U-shaped valley is representative of the many great fiords that trench the upland. The lower diagram is schematic for the Klamaths of northern California and for the coastland of this area. It should be considered as a north-south section in the Klamaths with the horizontal lines representing sea level at different times along the coast. The horizontal lines in both diagrams represent different sea levels. Sea level 1 was the base to which the high surface in both regions was graded. Sea level 2 was the one after the great emergence to which the deep gorges were eroded. Sea level 3 was the one after the great submergence to which the highest beaches now remaining were eroded. Sea level 4 is the present one after appreciable emergence. In British Columbia and southeastern Alaska this last emergence has only recovered 600 feet of the previous 1600 feet of submergence, whereas in nothern California the recovery has been almost complete. The original great uplift was caused undoubtedly by deep-seated crustal disturbances, but the later submergence and emergence were due to isostatic adjustments to the loading and unloading of the glaciers.

land and dissected by streams flowing over it. In the lower diagram, Fig. 29.12, the horizontal datum line marked 2 indicates sea level at this time. The uplift was probably over 2000 feet in the Klamath area. Then followed a subsidence of over 1500 feet. Datum line 3 indicates the sea level at this stage. The oldest beaches known in the region were established at this time. The highest are 1500 feet above the present sea level. They remain only in remnants today. The deep and narrow valleys cut in stage 2 were partly alluviated in stage 3. Through a succession of uplifts, beaches were formed at successive levels down to the present, with the modern coastal plain not far above sea level as the last major beach.

Northward from the Klamaths in southern Oregon, the shore terraces gradually disappear. The same is true southward in northern California. The most recent submergence north of the 40th parallel can be detected in the tidal portions of the rivers which are somewhat drowned. The subsidence increases as far north as the Columbia.

These very considerable epeirogenic movements in late Pleistocene and Recent time must be viewed with respect, when the offshore submarine topography is considered, because they show how possible it is for ex-

tensive parts of the continental shelves to have been emergent and how quickly the geography can change.

SAN ANDREAS FAULT SYSTEM

Aspects of Controversy

Perhaps the most discussed and widely known structural feature of the western United States is the San Andreas fault. See index map, Fig. 29.2, for location. It may be traced with ease and certainty from Tomales Bay, 40 miles northwest of San Francisco, to Cajon Pass, 50 miles east of Los Angeles. It has also been traced with a little doubt and difficulty for some scores of miles northwest and southeast of these limits. Its total known length is, therefore, more than 600 miles. This fault is so conspicuous that it was well known even before April 18, 1908. On that date, it was the site of a violent earthquake in the vicinity of San Francisco.

There is much conflicting literature written about the age of the San Andreas fault, its movement, and its relation to the compressional folds and faults. Some believe it came into existence first in pre-Cretaceous time and moved recurrently through the Cenozoic to the present. Some view the movement to have been mostly vertical, others mostly horizontal. The vertical movement is said to be great, around 20,000 feet by some; and only a few feet, by others. Those who recognize horizontal movement are divided in their opinions. Some think the movement has been a few thousand feet, others 300 miles or more. The most perplexing problem about the San Andreas fault in the central ranges is its setting in typical compressional structures running parallel or at an acute angle to it. The great fault seems at odds with the geomorphic provinces.

Those who have studied the fault north of the Garlock fault commonly interpret it differently from those who have studied it southward. Dibblee, however, who has studied the fault system both north and south of the Garlock fault probably more extensively than any other geologist, sees right-lateral movement predominantly throughout the entire length (Hill and Dibblee, 1953).

Main Faults and Relations of the System

The master fault of the system is considered the San Andreas, and the Big Pine and Garlock faults principal conjugate sheers (Hill and Dibblee, 1953). See Fig. 29.2.

In the San Francisco Bay area the Hayward fault passes through Berkeley and the site of the University of California stadium. A little to the east is the parallel Calaveras fault. Branches of the San Andreas extend up the peninsula on the west side of the bay. No long faults have been mapped in the northern Coast Ranges except some just north of San Francisco Bay.

The Garlock fault is conspicuous from its position at the boundary of a region of strong relief on the north and subdued relief on the south in the Mojave Desert.

The San Jacinto and Elsinor faults are major ones in the Peninsular Ranges and most probably shared the horizontal movement with the San Andreas. In fact, most all the faults shown on the map of Fig. 29.2 are large, and probably parts of the system.

In studying displacements and ages of the faults the following rock types, as far as manner of response to deformation, have been distinguished (Hill and Dibblee, 1953):

1. Sierran basement complex (pre-Cretaceous): metasedimentary and metavolcanic rocks, intensely deformed and widely invaded by granitic rocks. Because of physical similarity, the Santa Lucia granitics and metamorphics of the southern Coast Ranges and the complexes of the Transverse and Peninsular ranges belong in this group. These are relatively rigid rocks which fail locally by fracturing and, since they or rocks like them are extensively exposed and are presumably of state-wide occurrence at depth, their mechanical behavior is tectonically important.

2. Franciscan basement (pre-Cretaceous): sedimentary and volcanic rocks, regionally unmetamorphosed but highly indurated, commonly intruded by basic igneous rocks which are usually altered to serpentine and have caused local metamorphism. These rocks are exposed in large areas in the Coast Ranges; on the northeast side of the San Andreas fault, and also on the west side of the Nacimiento fault zone. They presumably underlie a much greater area but are probably in turn underlain by granitic rocks. The Franciscan, unlike the granitic basement, is typically incompetent. Although in places intensely fractured, often before being covered by later Jurassic or Cretaceous strata, and usually in

fault contact with the other principal rock types, its response to deformational forces has been characterized by folding.

3. Cretaceous and Cenozoic sedimentary and volcanic formations: mainly marine clastic sediments with local volcanics and nonmarine deposits, not strongly lithified and of extremely variable thicknesses and facies. Deposited in large and small basins; locally highly deformed, especially during the late Pliocene–Pleistocene revolution in the Coast and Transverse Ranges, and in uplifts in the Mojave Desert and Salton Sea region regions. These rocks form a pliable mantle on the above described complexes and have therefore responded to tectonic forces primarily by folding, particularly where the sedimentary section is thick or where underlain by Franciscan basement.

The San Andreas fault marks such an important contact that rarely can it be crossed, except in Recent alluvium, without passing into significantly different rocks. It is also a steep, if not nearly vertical fault and extends to depths of at least 10 miles, according to seismological evidence.

Evidence of Horizontal Displacement

The following evidence of horizontal movement on the San Andreas fault is presented by Hill and Dibblee (1953):

1. The trace of the San Andreas zone is typically continuous and straight. There is evidence of recent activity along its entire course. Excepting a 30-mile segment trending eastward in the San Emigdio Mountains, and another stretch of similar trend 100 miles to the southeast, the zone is remarkably straight from Point Arena southeastward nearly to Mexico. These aspects of continuity and straightness are considered typical of strike-slip faults.

2. The San Andreas is a steep fault which transects major topographic features but develops all along its course one or several parallel trenches, sag ponds, low ridges, saddles, and/or scarps. Its steepness is indicated by the straight trace, the fact that mapped fault planes are nearly vertical, and the failure of near-by drill holes to penetrate the zone. These characteristics are typical of strike-slip faults. The development of fresh topographic features, many of which are in unconsolidated recent sediments, and the common lack of appreciable vertical or consistent vertical components of offset clearly indicate the recency of lateral movements. Seismic evidence for recent right lateral movements on the San Andreas, as summarized by Wallace (1949), comprises the following maximum displacements at the time of earthquakes: 30 feet (San Emigdio Mountains, 1857), 10 feet (San Francisco area, 1868), 21 feet (San Francisco area, 1906), and 10 feet (Salton Sea area, 1940).

3. The San Andreas fault zone ranges from a few feet to a few miles in width. Locally a single recent trace may be irregular, with 15-degree variations in strike within a few hundred feet, or it may disappear and be replaced, *en echelon*, by another. Occasionally two or three parallel traces widen the zone of recent traces to a maximum of about half a mile. Wider segments of the zone consist of several faults (not necessarily active) which are usually steep and nearly parallel to the trend of the zone. These characteristics are considered typical of strike-slip fault zones along which recurring movements have taken place.

4. The apparent throw is commonly reversed along the San Andreas fault as indicated by topographic and geologic relationships. These throws are probably due to the major strike-slip component which places in juxtaposition unlike topographic elevations and geologic sections, and thus the reversals of dip-slip are mainly illusory.

5. Drainage lines are consistently offset in a right lateral sense. These offsets are especially clear on the southwest side of the Temblor Range where a maximum of 3000 feet of displacement has occurred through recent movements on the fault. Wallace (1949, p. 805) reports a probable drainage offset of 1½ miles on the north side of the San Gabriel Mountains, and Allen (1946, p. 50) reports 3800-foot offsets of drainage lines near the Gabilan Range, also in a right lateral sense.

6. Recently developed trenches which trend southward into the fault have been observed in aerial reconnaissance on the southwest side of the Temblor Range. These are oriented correctly to be tensional in origin and due to right lateral movement on the San Andreas.

7. Locally developed west-northwest trending folds adjacent to the San Andreas are obviously drag folds resulting from the right lateral movement on the San Andreas. Such drag folds are expecially clear in the Salton Sea Region, and, besides indicating the right lateral sense of movement on the fault, many of them show by their discordance with topographic form that the fault was active before the present physiographic features were developed.

8. Wallace (1949) reports a probable 6-mile right lateral offset of terrace deposits on the north side of the San Gabriel Mountains, and L. F. Noble (personal communication) describes similar late offsets in that area of several miles.

9. Between the San Emigdio Mountains and the Temblor Range, there are two facies of Pleistocene gravels. On the southwest side of the San Andreas, the pebbles are granite, gneiss, quartzite, limestone, black shale, and sandstone which undoubtedly came from the San Emigdio Mountains. On the other side of the fault, the pebbles are almost exclusively white siliceous shale which probably came from the Miocene shale of the Temblor Range. These two facies are in direct contact along the San Andreas for several miles. Furthermore, the northwest end of the crystalline clast facies is about 14 miles northwest of the crystalline rocks of the San Emigdio Mountains. These relationships, thus indicate a right lateral displacement of approximately 10 miles on the San Andreas fault since Pleistocene deposition in this area.

10. In the Caliente Range, marine sediments of upper and middle Miocene age grade laterally eastward into continental red beds which strike into the San Andreas fault, whereas strata of the same age are marine shales on the

other side of the fault. This juxtaposition of unlike facies again demonstrates substantial lateral movement. In this case the general trend of the western margin of the continental facies in the Caliente Range is northward across the Carrizo Plain toward the San Andreas, whereas possibly the same transition line may be extrapolated southward from along the east side of the San Joaquin Valley to the fault. Thus, by simple projections the right lateral offset on the fault since the upper Miocene time would be about 65 miles, although the probability of irregularities in trend of this facies contact precludes a strictly quantitative solution of that cumulative shift. Note the comparable offset of the upper Miocene "Pancho Rico"–"Santa Margarita" shale, shown in the same figure.

11. Going back only slightly farther in the geologic record, approximately 175 miles of right lateral offset may have accumulated on the San Andreas fault since early Miocene time. This is suggested by the unique similarities of rock types and sequences in the San Emigdio Mountains, as described by Wagner and Schilling (1923), and the Gabilan Range as described by Kerr and Schenck (1925), and Allen (1946). In each of these areas, a section of lower Miocene volcanics, red beds, and marine lower Miocene and Oligocene strata occurs [B–B′ of Fig. 29.13].

12. A similar relationship is suggested by some lithologic and faunal similarities between the Eocene formations of the Temblor–San Emigdio and the Santa Cruz Mountains which indicate the possibility of an offset of approximately 225 miles since late Eocene time [C–C′ of Fig. 29.13].

13. Also the southern limit of Cretaceous strata in the Temblor Range may match with the southern limit of Cretaceous beds near Fort Ross which would indicate an offset of approximately 320 miles [D–D′ of Fig. 29.13].

These evidences of progressive movement from the Cretaceous to the present are consistent with each other and yield a rate of 0.2 to 0.3 inch of movement per year. However, geodetic measurements of rates since the turn of the century are about tenfold the ones based on offsets of rock masses (Hill and Dibblee, 1953).

Contrary to the substantial evidence of large horizontal movement south of San Francisco, Higgins (1961) concludes that less than 15 miles of right-lateral displacement has occurred along the San Andreas north of San Francisco since mid-Pliocene time. During the same time the east side has been raised about 500 feet relative to the west side.

Big Pine and Garlock faults

Both the Big Pine and Garlock faults have left lateral movement in contrast to the right lateral movement of the San Andreas. The one is

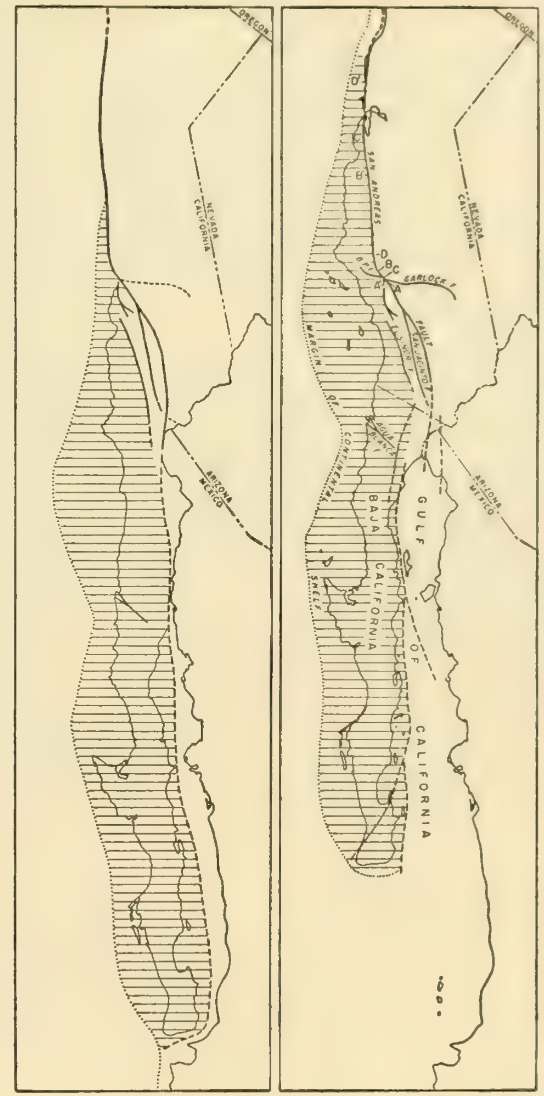


Fig. 29.13. Maps showing postulated strike-slip movement along the San Andreas fault. Left map shows position of Baja California and Coast Ranges of California (shaded area) in Cretaceous time. Right map shows the present position. D and D′ were juxtaposed in Cretaceous time; C and C′ in Eocene; B and B′ in Oligocene and early Miocene; D and D′ offset of Big Pine fault. Hill and Dibblee, 1953; Hill, 1954.

believed to be the offset of the other (Hill and Dibblee, 1953). The shift to the northwest has been about 5 miles. See A–A′, Fig. 29-13. If such is true, and if the horizontal displacement along the San Andreas has been in the order of 300 miles, then the San Andreas is much older and had been active a long time before the Big Pine–Garlock fault came into existence. The 5 mile offset would indicate that the age is Pleistocene. There is no question about the recency of activity along the Big Pine and Garlock faults, but the time of beginning may be suspect. It could be that the Big Pine fault originated many miles to the south and by coincidence is now about opposite the Garlock.

Hill and Dibblee believe the strain system of the Big Pine–Garlock shear and the San Andreas shear is a conjugate or complementary one with the south wedge moving against the north wedge. Moody and Hill (1956) elaborate on the stress-strain relations of the San Andreas system in which they call the strike-slip faults of large displacement "wrench faults." They develop second and third order effects and believe they demonstrate at least eight directions of wrench faulting and four directions of folding or thrusting possible. They conclude that dynamically the orientation of the Garlock is not correct for the primary left lateral direction, and it would more nearly fit a theoretical position for a second-order left lateral fault, assuming north-south compression. The Transverse Range may represent the primary fold direction, consequently shortening the crust in this area and altering the San Andreas direction (Moody and Hill, 1956).

Relation to Pacific Fracture Zones

Great fracture zones trend generally westward from the United States, Mexico, and Central America across the Pacific. These are depicted in Chapter 32, and their relation to the fault system of California is shown in Fig. 32.15. The relation of the two systems is an enigma.

Origin of Gulf of California

Recent seismic work in the Gulf of California has shown the crust there to be oceanic (H. W. Menard, personal communication), and consequently one's impulse is to postulate drift of the peninsula away from the mainland. Not only westward but northwestward drift compatible with movement along the San Andreas fault must be postulated. If the Coast Ranges oceanward of the San Andreas fault and the Peninsular Ranges with Baja California are moved as a unit southeastward in the amount of movement proposed by Hill and Dibblee from the Cretaceous to the present, the Baja California is brought into a likely former position with the mainland. The Nevadan belt of the Sierra Madre del Sur would continue in this arrangement without break into Baja California, as postulated in Chapter 38. The restored relations are shown in Fig. 29.13.

Two difficulties appear; the long unit has to be bent slightly to make the fit, and it has to snake around the major bend of the San Andreas fault east of the Los Angeles Basin in making its way to the northwest. The passage is accomplished in a more straight-away course if a good deal of the movement occurred along the Elsinor, San Jacinto, and San Gabriel faults. Hill and Dibblee have commented that the San Gabriel fault may have been principally active in the past. It seems possible that the segment of the strip now making up southern California has been pressed somewhat against the continent since late Miocene time, and although right lateral movement has continued along the San Andreas fault that the folds and thrusts of the Transverse Ranges were thereby formed. If a subcrustal convection current is carrying the strip northwestward, then the current might have become a little deflected toward the continent in the southern California region and the unusual complex of structures formed there. We could imagine that the Elsinor fault first carried the brunt of the dislocation, then the Jacinto, and finally the San Andreas through the Salton Sea area, as the compressive component of the carrying force increased against the continent. All these faults are still comparatively active.

Seismicity in the Coast Ranges

Figure 29.14 shows the general seismicity of the California region. The epicenters are scattered through the San Andreas fault zone more widely than might be expected, yet there is a general clustering along the great fault. The Agua Blanka fault of northern Baja California is believed to

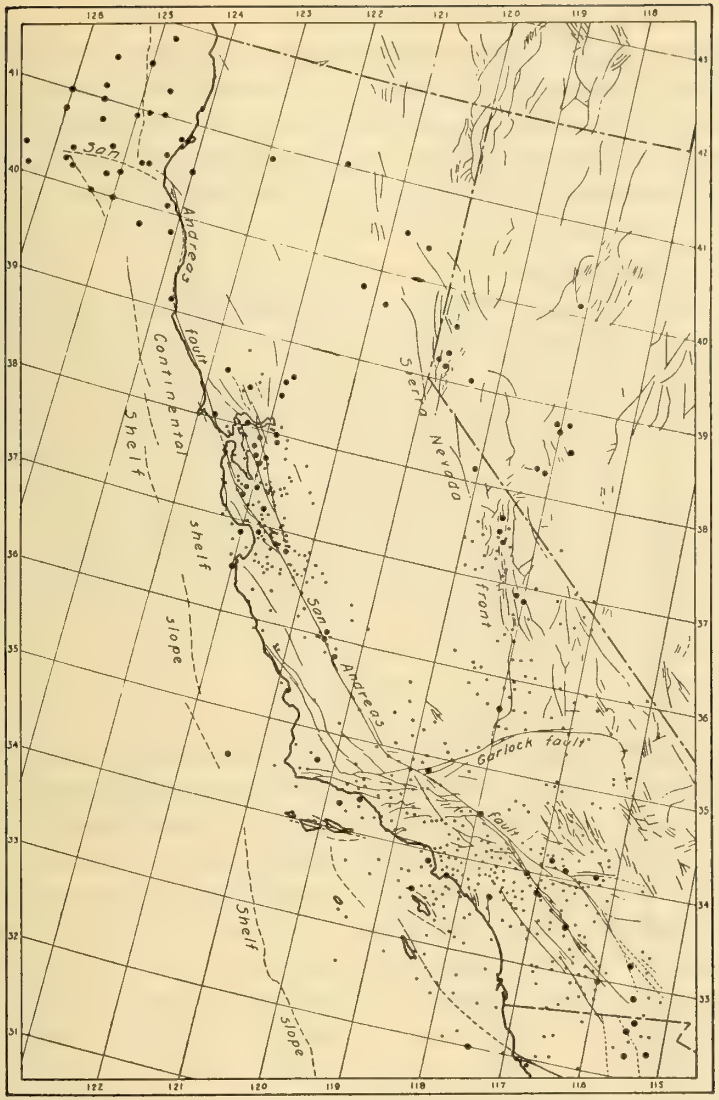


Fig. 29.14. Earthquake shocks and faults of California and western Nevada. The faults are those generally considered to have suffered late Pleistocene or Recent activity. Earthquakes compiled from Byerly (1940), Gutenberg (1941), Byerly and Wilson (1936, 1937), and tables supplied by C. F. Richter. Earthquakes above the magntiude of 5 are shown in large dots, those below by small. In the compilations some earthquakes may have been shown twice because of overlapping and discrepancies in location assignments.

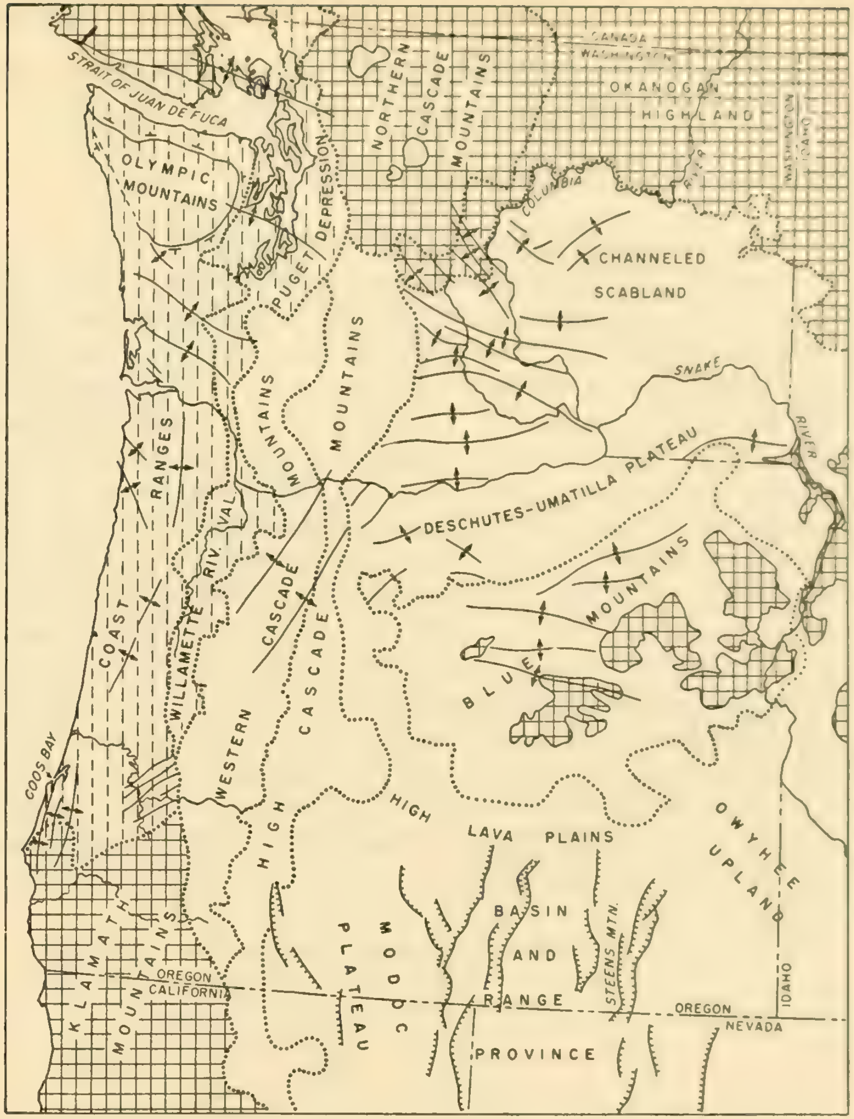


Fig. 29.15. Index map of Washington and Oregon. Coast Ranges are vertically dashed; the extensive volcanic fields are unruled; and the pre-Tertiary rocks, mostly Nevadan complex are cross-ruled.

project out to the San Clement basin and escarpment on the basis of the epicenters and submarine topography (Allen *et al.*, 1960).

COAST RANGES OF OREGON AND WASHINGTON

Geomorphic and Geologic Provinces of Oregon and Washington

Figure 29.15 has been prepared to show the geomorphic provinces of Oregon and Washington, and in a broad way the geologic divisions. The Klamath, Blue, and Northern Cascade Mountains, and the Okanogan Highland have been referred to in Chapters 6 and 17. They are made up chiefly of the Nevadan complex. The trends in the Klamath Mountains veer northeastward as they pass under the Tertiary volcanics and are generally thought to find a continuation in the Blue Mountains. Most of the sedimentary rocks of the Blue Mountains are unmetamorphosed, and this is puzzling because the rocks of the Nevadan complex elsewhere are fairly crystalline. The large Idaho batholith lies east of the Blue Mountains and appears to make up a knot at the intersection of the Sierra Nevada–Klamath–Blue arc and the British Columbian Coast Range arc with its great batholiths. The basement geology of Oregon and Washington is thus believed to be the Nevadan complex at the junction region of two great arcs. It evolved as a Paleozoic and early Mesozoic eugeosyncline. In Late Jurassic and Mid-Cretaceous time folding, metamorphism, and batholithic intrusions brought its history to a climax. The Tertiary Coast Ranges and the extensive volcanic fields developed thereafter.

As in California the Coast Ranges are bordered on the east by a general depression, known in Oregon as the Willamette River Valley, and in Washington as Puget Sound. The two are referred to as the Willamette–Puget depression or Willamette–Puget Sound depression. On the east of the depression are the Cascade Mountains, made up of volcanic rocks. They are divided into the Western Cascades and the High Cascades as shown in Fig. 29.15, and are treated fairly extensively in Chapter 36.

East of the Cascade Mountains and surrounding the islands of pre-Tertiary rocks in the Blue Mountains are vast Tertiary volcanic fields. North of the Blue Mountains and including part of them is the Columbia River basalt field, and south of the Blues are several geomorphic provinces all underlain by volcanics, sometimes collectively referred to as the Malheur field. The southern lavas are generally younger than the northern. The Columbia and Malheur fields are outlined in Chapter 33.

Divisions of Coast Ranges

The Coast Ranges of Oregon and Washington are a coherent unit geologically, because their formations are probably all Tertiary and they have been deformed as a unit. The northern end is composed of the Olympic Mountains, a domal uplift supporting the highest peaks of the Coast Ranges, with Mount Olympus 7954 feet above sea level. The canyons of the Olympic Mountains have been heavily glaciated.

At the northern end of the Coast Ranges of Oregon, just south of the Columbia River and west of the city of Portland is another uplift in which a core of fairly old rocks (middle Eocene) relative to those of the ranges elsewhere is exposed.

Stratigraphy

Selected sections of the Cenozoic rocks of the Coast Ranges of Oregon and Washington are given in Fig. 29.16. They are taken from Weaver's (1945a,b) extensive study with the western Oregon section modified according to Baldwin (1959) and Wilkinson (1959). The idealized cross sections, A–A′ and B–B′ of Fig. 29.17, attempt to restore the deposits to their condition before the late Miocene folding.

At the beginning of Tertiary time, according to Weaver (1945), a vast erosion surface existed in eastern and western Washington in the manner of a coastal plain. It had been carved chiefly in the rocks of the Nevadan orogenic belt. Early in the Eocene, the plain began to subside, and the earliest deposits filled the broad valleys of the extensive erosion surface. The Swauk formation of eastern Washington may be a fresh-water deposit in the upper part of one of these valleys, and the Solduc formation of the Olympic Mountains may be the marine equivalent. Both of these formations were folded somewhat and eroded before the overlying vol-

canics were poured out. These outpourings have been called the Teanaway volcanics in eastern Washington, the Metchosin volcanics in western Washington, and the Tillamook and Siletz volcanics in western Oregon.

The basal Eocene volcanics are a voluminous deposit. They originally formed a vast lava field that extended from Vancouver Island 500 miles southward to the Klamath Mountains and from a line considerably west of the present coast 150 miles inland. Their minimum average thickness was 3000 feet. According to Weaver, the volume of these volcanics was greater than the Columbia plateau basalts. They consist mainly of andesitic and basaltic flows with tuffs, agglomerates, and numerous intrusive plugs and dikes. The latter crosscutting intrusions, Weaver believes, were the vents of much of the volcanic material.

By the close of the Metchosin volcanism, a narrow north-south trough formed with its axis in the approximate position of the present Willamette–Puget Sound depression, and its sediments extended westward into the site of the modern Coast Ranges. After the volcanic eruptions 8000 to 14,000 feet of sediments were deposited. They make up the Puget group of the Seattle region, the Cowlitz formation southward in Washington, and the Tyee sandstone and Coaledo formation in Oregon.

The basal volcanics remained emergent in a narrow peninsula that projected southward from Vancouver, with the trough to the east. In early Oligocene time the peninsula submerged in part, and sediments were deposited directly on the Metchosin volcanics there; farther east they rest on the late Eocene strata of the trough. By late Oligocene, the peninsula area had sagged so much that 8000 feet of sandstone and shale had accumulated. Again in middle Miocene time, over 4000 feet of sandstone and shale, the Astoria formation, were deposited in the Coast Range area.

During Miocene times, great quantities of lavas were coming to the surface through numerous vents and fissures, especially in the areas of the Columbia plateau and the present Cascade Mountains. These flows fingered out westward, but north of Portland they are particularly abundant and form about 50 percent of the Astoria formation (Weaver, 1945). See cross sections A–A′ and B–B′, Fig. 29.17.

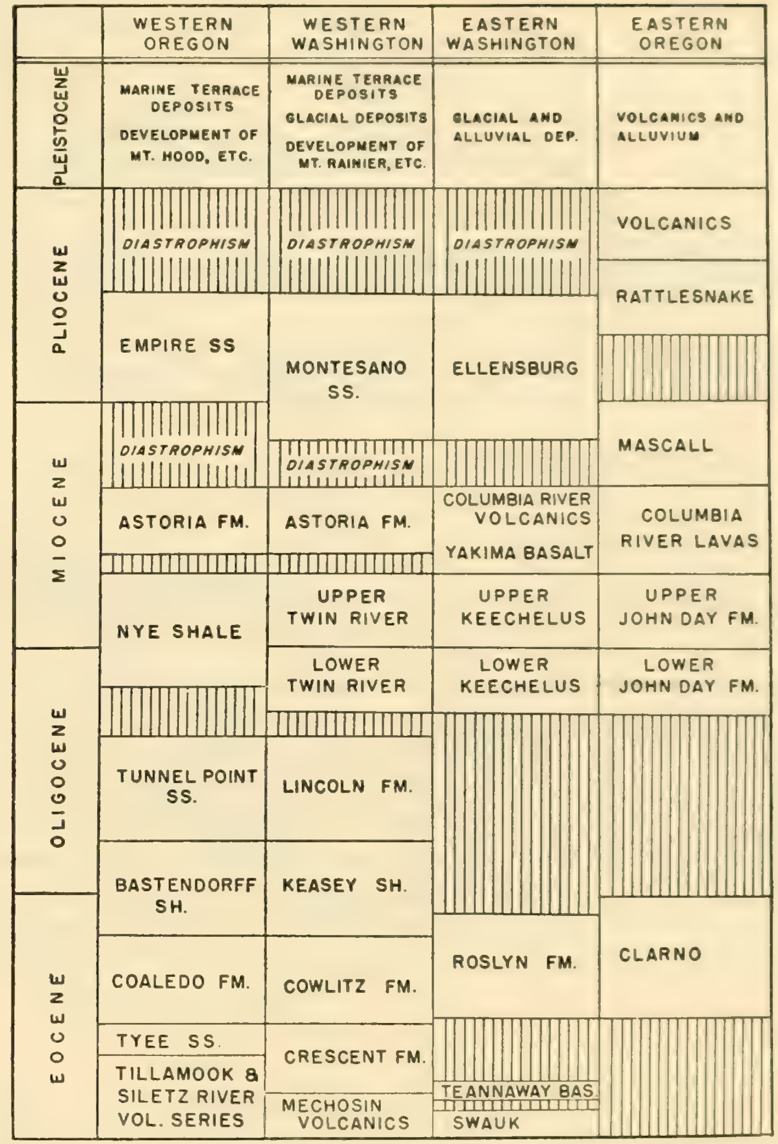


Fig. 29.16. Representative stratigraphic sections of the Tertiary in Washington and Oregon. After Weaver, 1944.

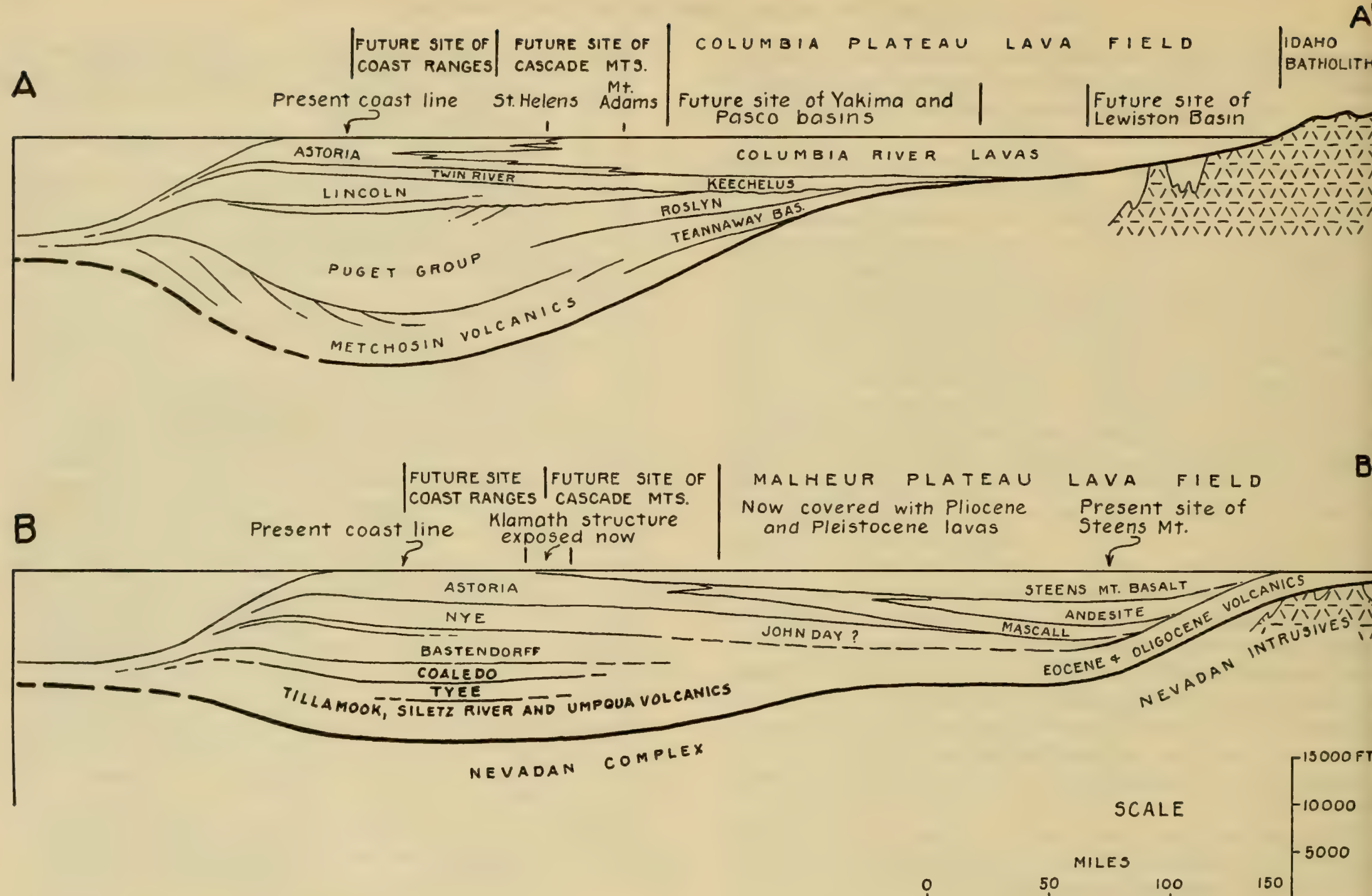
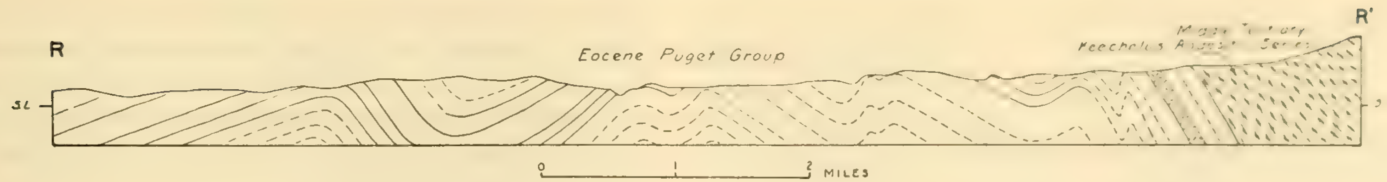


Fig. 29.17. A–A′, cross section through Washington and B–B′, cross section through Oregon. For positions see Fig. 29.1. They attempt to restore ideally the Eocene, Oligocene, Lower and Middle Miocene sediments and volcanics just before the folding in the trough area of late Miocene time.



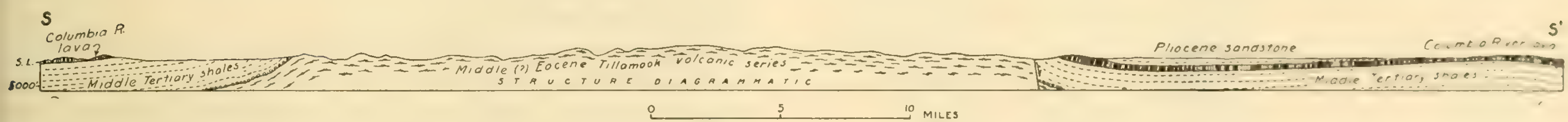


Fig. 29.18. Section R–R′ west of Tacoma in King County, Washington. See index map, Fig. 29.1. After Warren *et al.*, 1945a.

Section S–S′ across the Coast Range From Cape Meares to Williamette River, Oregon. After Warren *et al.*, 1945b. The Tillamook volcanic series is probably equivalent to the Metchosin volcanic series.

Late Miocene Phase

After the deposition of the Astoria formation, the trough sediments of Washington and Oregon were subjected to compression. As far as known, mostly open folds resulted. Perhaps in places they were compressed so as to have steep flanks or to be overturned. Examples are given in cross sections R–R′, and S–S′, Fig. 29.18. Faults are not common, and where present have been illustrated as the normal type. The fold axes that are known to have originated in this late Miocene phase have been assembled on the index map of Fig. 29.15. Through Washington, according to Weaver, they pass in a west-northwest direction. The axes that Weaver shows are those of very broad folds defined by the Vancouver Island–San Juan Islands–northern Cascade upwarp and the Olympic–Newcastle Hills–Cascade upwarp, with the intervening downwarp of the Strait of Juan de Fuca. Also, the Columbia River lavas in the western part of the basalt basin have been deformed into several northwest-trending broad anticlines and synclines. A map by Warren *et al.* (1945b) just west of Puget Sound (locality of section R–R′, Fig. 29.18) shows the folds to be small and rather tight, and they curve sharply from a west-northwest direction to a southerly and southwesterly one. The area covered by the new map is so small, however, in relation to that of the state and the broader picture, that the significance of the local variations is not known. The Miocene folds of the state seem to be of low to medium intensity and to trend generally to the northwest.

In the Portland area of the Coast Range, the fold axes are gentle and also extend in a northwest direction. They show a tendency to bend southward and generally parallel the coast. Farther south in Oregon, they parallel the coast, and some even trend to the southwest in the northern Klamath Mountains.

The Olympic Mountains uplift is ringed by a horseshoe-shaped exposure of the Metchosin volcanics with the Solduc formation underneath and presumably forming the core. The latter is more metamorphosed than the Metchosin and consists of phyllites and argillites. It seems to have great thickness. However, Oligocene fossils have been found in the area of Solduc (?) rocks, and thus the simple dome structure is doubted. Park (1950) concludes that the uplift contains steeply dipping thrust faults, and considerable buckling, thus reducing the previous estimates of a very great thickness for the Solduc.

Late Pliocene and Early Pleistocene Phases

Deformation at the close of the Tertiary and in the Pleistocene throughout the Oregon and Washington region has been of the broad arching,

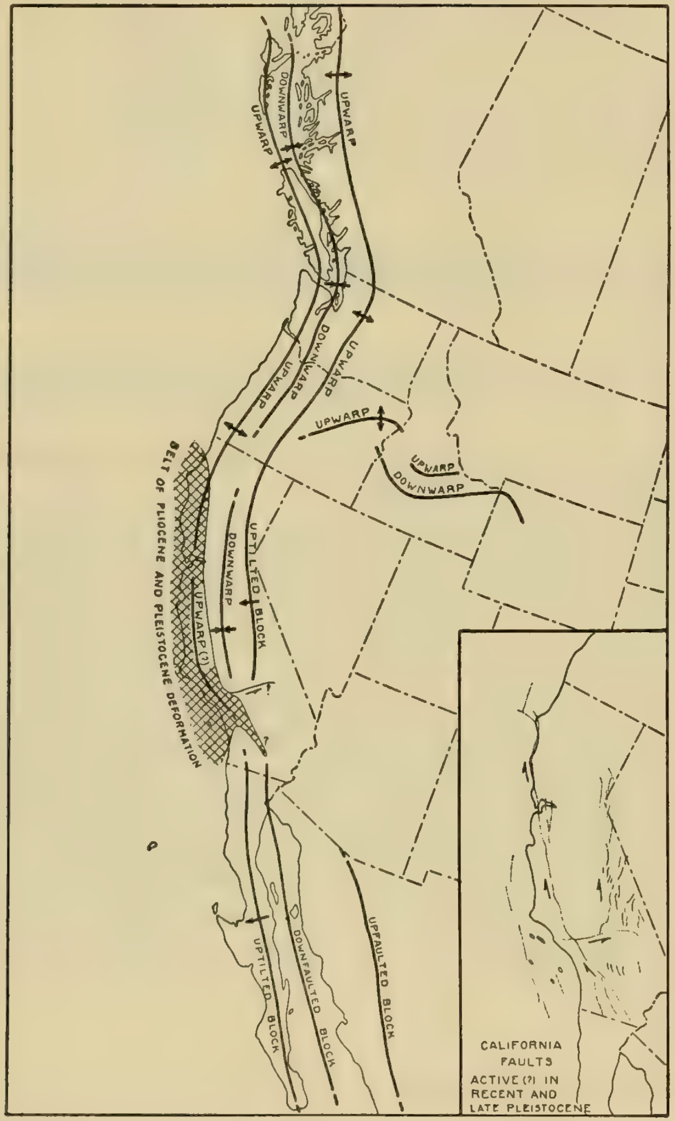


Fig. 29.19. Tectonic map of the late Pliocene and Quaternary crustal movements along the Pacific.

sagging, and warping type, and therefore contrasts sharply with the close folding, thrusting, and wrench faulting in the California Coast Ranges. The main orogeny of the southern ranges as previously pointed out occurred in late Pliocene and mid-Pleistocene time, but it mostly escaped the Washington and Oregon ranges. On the other hand, the late Miocene deformation seems to have been about of equal intensity both north and south of the Klamaths.

The gentle archings have been deduced from several lines of evidence. The first and most conspicuous is the parallelism of the three major topographic features, namely, the Cascade Range, the Willamette–Puget Sound depression, and the Coast Range. The two ranges are taken as arches or broad, gentle anticlines, and the depression as an intervening broad, gentle syncline. The second line of evidence comes from erosion surfaces, both inland and coastal. The third concerns the glacial deposits, which are very extensive in parts of the Puget Sound depression, and vertical crustal movements associated with the glaciation.

According to Weaver (1931–37) the Pliocene deposits, where known, rest unconformably upon the Miocene and are much less tilted. Thus the late Miocene phase is dated. During the latest Miocene and Pliocene, minor differential movements allowed the oceanic waters to transgress easterly and cover small restricted areas on the western side of the Olympic peninsula and in the coastal portion of southwestern Oregon. All other areas were undergoing erosion, and it is probable that the major channels of Puget Sound, such as Hood Canal, Admiralty Inlet, Georgia Strait, and the Strait of Juan de Fuca, were being excavated. The marine waters that occupy these valleys at the present time gained access as the result of Pleistocene depression just preceding and during the glacial epoch.

Near the close of Pliocene time the two broad anticlines and intervening syncline developed and emphasized the individuality of the Coast and Cascade ranges and the Puget trough (Weaver, 1937). See map, Fig. 29.19. These north-south structures were probably superposed on the Miocene northwest trending folds. The Cascade Mountains ultimately attained their present elevation during the early Pleistocene, and upon their surface was built a row of majestic volcanic cones such as Mount

Baker, Mount Rainier, Mount Adams, Mount St. Helens, Mount Hood, and numerous smaller cones in southern Oregon. See Chapter 33.

It seems probable that the erosion surface, developed after the late Miocene folding, was itself gently folded, as were the rocks beneath in the late Pliocene archings, and that it was intensely dissected where uplifted most.

After the elevation of the erosion surface, and after its deep dissection by the voluminous streams of the region, the ice age came on, and is recognized in two stages. During the later advance all the valleys of both the eastern and western slopes of the Cascade Range were filled with ice, which moved downward to lower elevations and built terminal moraines. The valley glaciers in northern Washington entered the Puget Sound basin and coalesced with one another, and with the extensive piedmont glacier that had moved southerly between Vancouver Island and the mainland. This great ice floe broke into two tongues; one extended westerly through the trough of the Strait of Juan de Fuca, and the other moved southward into the southern part of the Puget Sound basin, where it built up a terminal moraine from the southeast corner of the Olympic Mountains easterly to the Cascades.

After the withdrawal of the ice, the crust has risen in the Puget Sound basin and along the coast of Washington and Oregon from 20 to over 200 feet. A most recent submergence has already been noted near the mouth of the Columbia River, and the tidal influence extends eastward to the Cascades.

30.

BAJA CALIFORNIA AND SONORA SYSTEMS

BAJA CALIFORNIA

Topography

Baja California is as long as California but only a third as wide. See maps of Figs. 28.1 and 30.1. Its northern half is mountainous with peaks that rise to elevations of over 10,000 feet. These comprise the Peninsular Range, which is a continuation of the ranges of southern California west of the Salton basin. Most of the high area is granite and metamorphosed rocks of the Nevadan type. The southern half of the peninsula is lower in relief and for the most part is a great area of conglomerates, sandstones, agglomerates, and lava flows of post-early or middle Miocene age. See

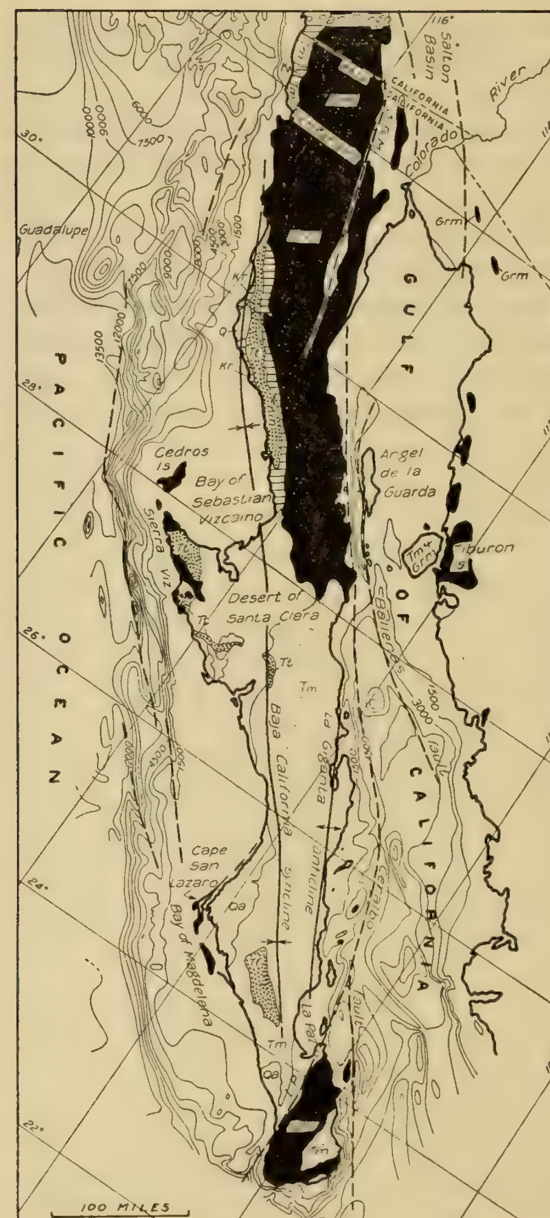


Fig. 30.1. Geologic and tectonic map of Baja California and the Gulf of California, after Beal, 1948. Grm, crystalline rocks of Nevadan complex; Km, Lower Cretaceous San Fernando formation; Kr, Upper Cretaceous Rosaria formation; Tt, Paleocene or Eocene Tapetate formation; Tm, Oliogocene to Pliocene formations.

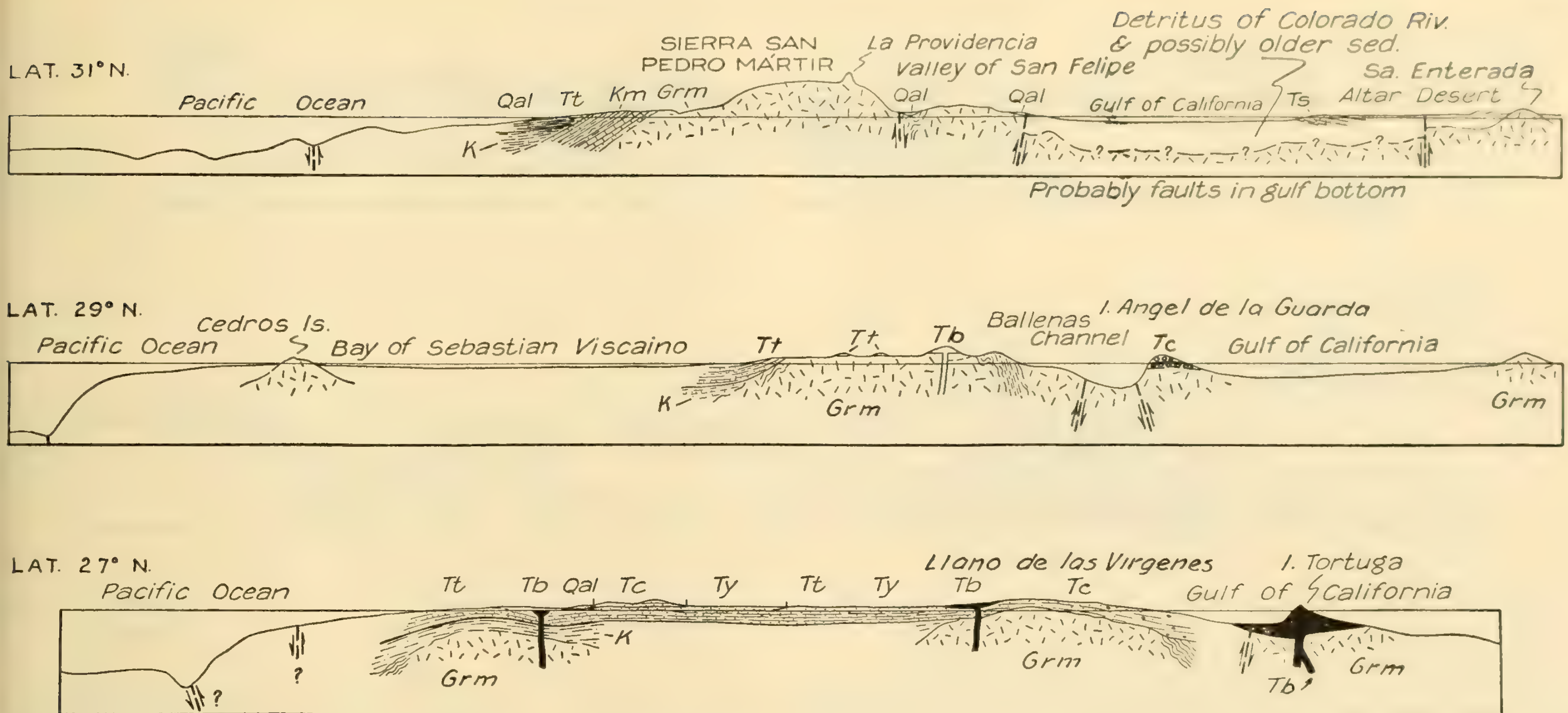


Fig. 30.2. Cross sections of Baja California, after Beal, 1948. Positions designated by latitudes are approximate.

Grm, crystalline rocks of Nevadan orogeny, diorite varieties, schists, gneisses; Tb, Cenozoic flows and intrusions, mostly andesite and basalt; Km, San Fernando formation of Cretaceous metamorphics, limestone, shale, quartzite and intrusive rocks; K, Rosario Upper Cretaceous marine sandstone and dark shales; Tt, Tepetate silts, sandstones, Paleocene to Eocene; Ty, Ysidro sandstone, siltstone and tuffs, Miocene; Tc, Comondú volcanics and clastics, upper Miocene; Ts, Salada formation, Pliocene.

cross sections of Figs. 30.2 to 30.4. Inland southeastward from the Bay of Sebastian Vizcaino is a vast desert of Quaternary alluvium about 800 feet above sea level.

The Tertiary rocks are in two narrow belts along each side of the Nevadan core in the northern part. Just south of latitude 30°, however, the Pacific belt of Tertiary deposition (perhaps only the conglomerates and volcanics of the upper Miocene) extends across to the east coast. Then from latitude 29° southward to the southern end of the peninsula, the Nevadan rocks crop out in the bordering islands on both sides, including the western cape region (Sierra Vizcaino). Finally, a large part of the south end of the peninsula (Sierra Victoria) is made up of the Nevadan complex. See the tectonic map, Fig. 30.1.

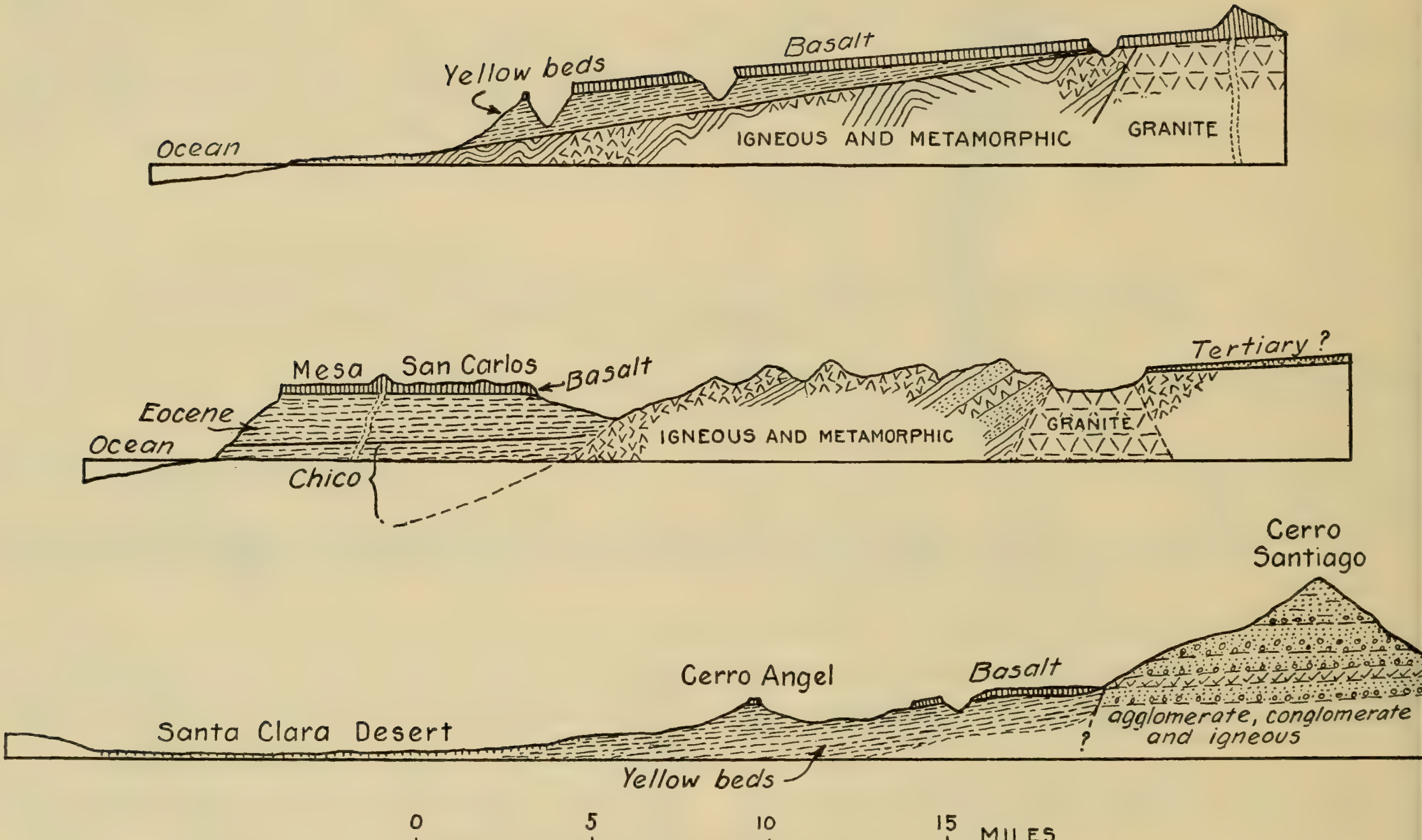


Fig. 30.3. Sections across central Baja California. After Darton, 1921.

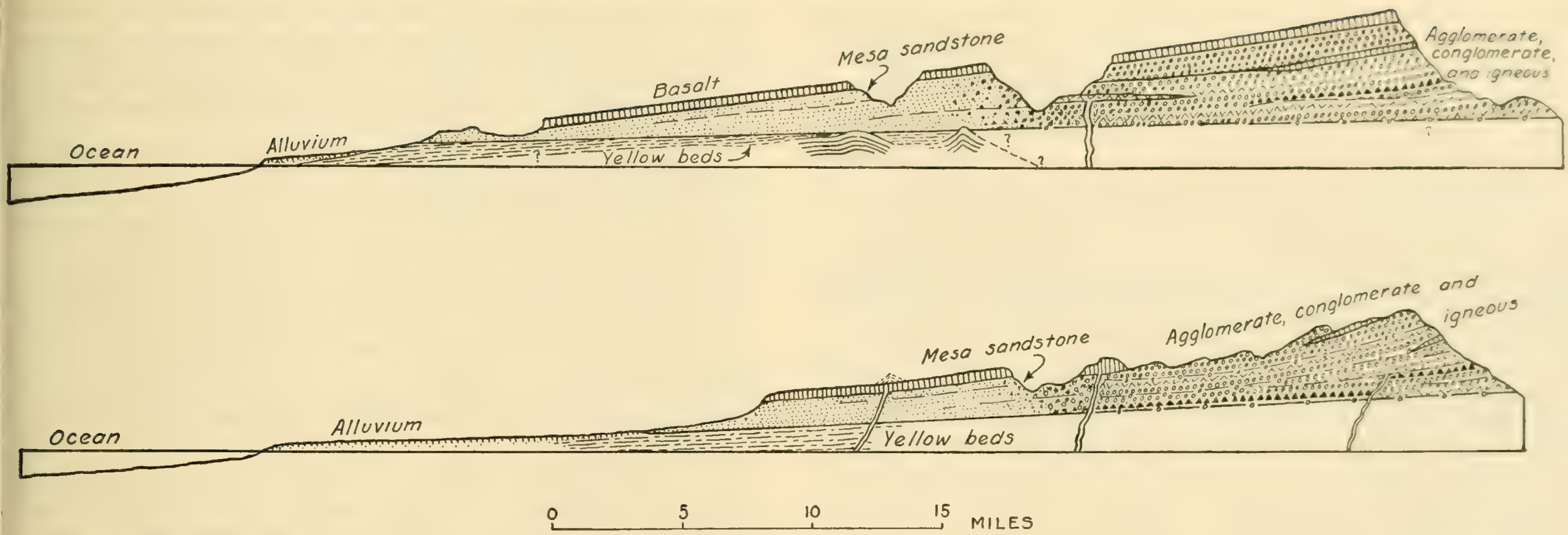


Fig. 30.4. Sections across parts of northern and central Baja California. After Darton, 1921.

Stratigraphy

Beal (1948) records:

The rocks of Baja California consist of (1) unaltered marine sedimentary rocks ranging in age from Cretaceous to Pleistocene; (2) a series of sedimentary rocks of probable Cretaceous age exhibiting varying degrees of alteration; (3) extrusive rocks, principally of andesite and basalt; and (4) intrusive rocks consisting principally of quartz diorite and granodiorite, which have intruded and metamorphosed older rocks, the age of which is not definitely known.

The Cretaceous is represented by the San Fernando and the Rosario formations. The San Fernando formation, Lower and early (?) Upper Cretaceous in age . . . consists of a series of slates, conglomerates, quartzite, limestone, and sandstone, with varying amounts of associated intrusive and extrusive rocks; some parts of the series are only slightly, but others greatly metamorphosed. The younger Rosario formation (Upper Cretaceous) is unconformable on the San Fernando formation. It consists of unaltered red and gray shale, brown sandstones, and conglomerates on the Pacific Coast near Rosario. . . .

The Tertiary is divided into the following formations: the Tepetate (Paleocene to Eocene in age) . . . composed generally of yellow to brown silt and sandstone; the San Gregorio formation, Oligocene (?) to Lower (?) Miocene in age . . . resembling in some respects the Monterey shales of California; the Ysidro formation, late Lower or Middle Miocene, or both, in age, comprising a lower member of shales, in part diatomaceous, and an upper member of light-colored sandstone and shale; the Comondú formation, Upper Miocene (?) in age, composed mainly of agglomerates, tuffs, and lavas; and the Salada formation (Pliocene), and consisting mainly of yellow marine sandstone and shale.

Following are the outstanding geologic features in the different districts of Baja California:

1. The Northern district is characterized by a high, westward-sloping block of crystalline rocks, which appear to owe their elevation to profound faulting along the east side. The axial mountains have a granitic core, flanked on both sides by gneisses, schists, and slates probably of pre-Cretaceous age; the bedrock complex is overlain on the west side by irregularly metamorphosed rocks of Cretaceous age, which are themselves overlain by unmetamorphosed marine sediments of Cretaceous and Tertiary ages; these rocks do not appear in any outcrop of importance on the east side. The crystalline rocks are prominent but decrease in elevation as far south as the 28th parallel, the southern boundary of the Northern district.

2. The major feature of the Western Cape region (28° and 114°) is the northwesterly trending Sierra Vizcaino, bordered on the north and east by Desierto de Santa Clara. Crystalline rocks, including small areas of the Franciscan formation, constitute the bedrock complex of Sierra Vizcaino the islands farther northwest. Tertiary and probably some Cretaceous sediments, dipping in general northeastward toward the synclinal desert and southward toward the ocean, overlie this bedrock complex. Volcanism is a major feature of the southeastern part of this area.

3. The areal geology in the South-Central area is dominated by volcanic rocks of Tertiary age, which obscure some of the earlier marine formations, but where these formations are exposed, they usually occupy the axis of a syncline and a part of the sierran area, which is anticlinal. Crystalline rocks, probably elevated by faulting, are exposed on the southwest coast at Bahia de Magdalena, but only small areas of these rocks, at relatively low elevations, occur along the uplifted gulf coast of the peninsula.

4. The Southern Cape region (24° and 110°) is looked upon as a distinct structural block and is almost entirely granitic and metamorphic rocks, although some marine Tertiary sediments occur east of the high sierra.

5. The islands adjacent to the peninsula are composed principally of volcanic and granitic rocks.

Metamorphic and Intrusive Rocks

Regarding the metamorphic and intrusive rocks older than the Lower and early (?) Upper Cretaceous San Fernando formation, which itself in places is metamorphosed, the following passages are quoted from Beal (1948):

Lindgren (1889) states that the principal mass of the peninsula at 32° N. Lat. is an enormous granitic plateau with minor areas of highly metamorphosed and compressed slates, the granites appearing to be a "white hornblende granite similar to that of the Sierra Nevada of California." Emmons and Merrill (1894) in their examination of the area adjacent to the 30th parallel found rocks of the same type as those mentioned by Lindgren and to the eastward found metamorphic slates which led them to remark on the similarity of structural conditions and lithological character of the rocks in the two areas. According to Hirschi the granitic zone of the Northern district is flanked on the gulf side by old crystalline schists, of a sort not observed on the Pacific side; and, in the desert sierras west of the mouth of the Colorado, great and varied schist zones occur, which extend southeastward along the gulf coast almost to the 28th parallel.

The metamorphic rocks, mapped with the intrusive granitics, were observed during this study to consist of gneisses, slates, schists, and other metamorphics; they are exposed on both flanks of, and on, the batholith which makes up the axis of the northern part of the peninsula and are known farther north on both sides of the batholith in San Diego, Imperial, and Riverside counties, California. Lindgren (1888) states in referring to the slates on the west side of the range at the latitude of Ensenada that "one cannot fail to be impressed by the enormous extent of the granite and the small area occupied by metamorphic rocks. It seems evident that the slates are of but little depth and everywhere are resting, as detached fragments, on the granite."

. . . Woodford and Harriss (1938), in a careful study of the granitic and associated metamorphic rocks adjacent to 31° N. Lat., state that the crystalline rocks consist of stocks and batholithic masses of quartz diorite. They consider that the plutonic rocks in northwestern Baja California "are typically quartz diorite, as contrasted to the granodiorite or quartz monzonite, which is the commonest rock of the Sierra Nevada." . . .

. . . In parts of the Western Cape region the granitic rocks are greenish-gray diorite and pink granite, occasionally cut by large intrusions of serpentine. According to Hanna (1927), chert, presumably of Franciscan age, occurs on Isla San Roque, Isla Asunción, Islas San Benitos, and possibly on Isla San Geronimo, far to the north just below the 30th parallel; he also reports (1925, p. 268) "Franciscan cherts, sandstones, and conglomerate" on Isla Cedros. At Punta San Hipolito (on the south coast of Western Cape region) are quartzites, cherts, cherty limestones, and igneous rocks, which were mapped as the San Fernando formation but may be older. In this area, as well as near Punta Asunción, the granites and metamorphic rocks are intruded by dikes of serpentine, but they have failed to alter the near-by Eocene sandstones which usually dip toward the crystalline rocks, indicating that the serpentine dikes are older than the Eocene and that the Eocene has been brought to its present position with reference to the crystalline rocks by faulting. At Punta San Pablo, about 25 kilometers northwest of Punta Asunción (Scammon Lagoon quadrangle), Hirschi and De Quevain (1933) state that the greenish-black rocks of the high coast line are probably of amphibolite and gabbro (?) broadly intruded by pegmatites, and that on the south end of the "intensively folded Paleozoic range" of Isla Cedros they observed strongly altered, glaucophane-bearing diabase porphyritic dike rocks.

The same authors refer to a great peridotite intrusion at Cabo San Lazaro and Punta Entrada (Magdalena Bay quadrangle), which is shattered, penetrated by east-west dioritic or dioritic porphyry dikes, and usually wholly altered to serpentine. They refer to andesitic and basaltic rocks of Tertiary age, which overlie the basement complex exposed along Bahia de Magdelena. Lindgren (1889) states that Isla Santa Margarita is composed principally of crystalline schists, with some chloritic slaty rocks and talc and serpentine. The numerous reported and observed occurrences of rocks of Franciscan character constitute good reason to believe that the Franciscan formation of California extends southward as far as the Western Cape region and perhaps to Bahia de Magdelena.

The metamorphic rocks of the Nevadan complex of Baja California can be judged better by comparison with their northward continuations in

California. The southern California area has been summarized in Chapter 17 to which the reader is referred for details. In brief, Larsen believes that there are many bodies of metamorphosed rocks older than the granitic rocks. Originally the sediments were shales, impure shales, and sandstones.

The argillaceous metasediments are chiefly on the west side of the main batholithic masses and within them, and they are probably mostly Triassic in age. The quartzites and coarse sericite schists are on the east side and are probably Carboniferous in age. A body of mildly metamorphosed volcanics of Early Cretaceous age makes up part of the pre-intrusive complex on the west.

The batholiths and older metamorphic rocks are overlain by Upper Cretaceous strata, and the date of the main intrusion is some time within the Lower Cretaceous.

Structure

Nevadan System. The metamorphic rocks and granitic batholiths of the Nevadan system probably make up the basement complex the entire length of Baja California. The geologic map, Fig. 30.1, shows a single great batholith extending halfway down the peninsula to the Desert of Santa Clara, and numerous other granite bodies carry the Nevadan system southeastward and end in the large batholith of the southern cape region south of La Paz. Islands on both the east and west coasts are composed entirely or in part of Nevadan complex. The Nevadan complex has been described, so the following structural study will deal with the Cenozoic folds, faults, and uplift.

Anticlines and Synclines. Beal (1948) has mapped a long, gentle anticline and an almost equally long, gentle syncline in the southern half of the peninsula. See map, Fig. 30.1. The syncline, known as the Baja California, extends from 31° N. Lat. southward for 600 miles to the isthmus of La Paz.

For the first 200 kilometers of its course it follows the Pacific Coast, first offshore and then on land, with marine sediments dipping gently toward its axis. At 29° 30′ N. Lat. near Bahia San Carlos, the syncline leaves the peninsula and crosses Bahia Sebastian Vizcaino, enters the peninsula again in the northwestern part of Desierto de Santa Clara, and extends thence through the desert in nearly a straight line toward the Isthmus of La Paz. From the south part of the desert at 27° N. Lat. most of the marine sediments dip gently toward the axis of the syncline, but, throughout much of this segment, these sediments are overlain and piled high with Comondú and later volcanic debris. Numerous local folds, some gentle, but others sharp, narrow wrinkles, were found in the trough of this great syncline.

The anticline along the east coast is called La Gigante and extends from Santa Rosalia southeastward about 200 miles to the Bay of La Paz.

Detailed mapping will undoubtedly show the area through which the axial line is drawn to be generally anticlinal and will probably disclose that the uplift is made up of several discontinuous anticlines, and faulting has been a factor in its elevation. Over nearly the entire distance the elevation of the area has resulted in great coastal escarpments which rise steeply for hundreds of feet from the gulf shore or the narrow coastal plain.

Bahia Concepción is definitely anticlinal as the Comondú rocks on both sides dip gently in opposite directions, and the same rocks near Aguaje at the southern end of the bay are folded into a well-marked anticline and several smaller folds, with dips ranging up to 30°. North of Loreto (Comondú quadrangle) the mountain shown as 2227 feet high has been forced up causing the Pliocene sediments to dip away in all directions. This area lies on another anticline east of the major uplift, but a few kilometers southwest granitic rocks are exposed on the axis of La Gigante anticline with Comondú rocks arching over the exposure. Southeast of Agua Verde (Santa Cruz quadrangle) the mountains back of Punta San Marcial are distinctly anticlinal, and east of Rancho Carrizalito (Santa Catalina quadrangle) the Ysidro formation is arched over a small mass of crystalline rocks.

Faults. The great eastward-facing escarpment along the Sierras Juarez and San Pedro Mártir is believed to mark a fault zone which has been called the San Pedro Mártir by Beal (S.P.M. on map of Fig. 30.1). The scarp is likened to that of the Sierra Nevada, and the fault zone is thought to be continuous along the east side of the Peninsula Range into southern California. Beal (1948) judges the vertical displacement to be about 5000 feet at the 31st parallel.

The western face of the Sierra Victoria of the southern cape region is considerably steeper than the eastern, and is regarded by Beal as marking a fault, called the La Paz (L.P. on map, Fig. 30.1). Submarine contours northward in the gulf suggests a projection of the fault. Beal points out that the Sierra Victoria trends northward obliquely across the peninsula and stands apart as a distinct unit. It thus seems to require a structure

such as the postulated La Paz fault, which he considers pre-Tertiary in age.

The submarine topography of the Gulf of California will be described immediately, and a downfaulted origin postulated. The Ceralbo fault is the major dislocation visualized.

Darton (1921) believed the major orogeny of the peninsula in Tertiary time resulted in the tilting of the long block upward on the east side and the sinking of the gulf, as diagrammed in Fig. 30.5. This presumably is the overall picture, but Beal adds three other structural elements, namely, the long, gentle folds and the diagonal La Paz fault, the major faults indicated by the submarine topography along the west side of the peninsula, and regional uplifts and submergences in late Cenozoic time. The submarine topography is treated separately in Chapter 32 and the regional vertical movements in the following tectonic history.

Tectonic History

The following résumé of the tectonic history of Baja California is composed of quotations from Beal's (1948) memoir.

Cretaceous Phase. The earliest record of the Cretaceous in Baja California is the San Fernando formation, which, insofar as it is known, was deposited only on the western slopes of the peninsula. Its lower stratigraphic limit is not known. The area over which the formation occurs probably was subjected to erosion during a long period before the deposition of the Rosario sediments and was extensively intruded during that time, which in places almost obliterated the sedimentary character of the series. No intrusions of the same type were observed to cut the younger Rosario formation.

The base of the Rosario formation was not seen as it probably lies under the ocean, and the series may be much thicker than indicated by the exposures. . . . During the deposition of the Rosario formation considerable areas of the San Fernando formation and of the earlier metamorphic rocks stood above water; erosion, whether shoreline or by streams, was principally in such rocks; and parts of the Rosario formation were also above sea level while sediments of the same series were being deposited. . . .

Following the deposition of the Rosario formation, the strata were locally distorted, but where observed, they were not usually folded sharply nor faulted. These structural phenomena suggest compression and folding while the sediments were but slightly loaded and before they had been completely indurated. . . . The diastrophic activity resulted not only in the mild folding and partial erosion of the Rosario sediments, but also marked an important emergence extending as far south along the west coast as 28° N. Lat., because the succeeding Eocene beds north of that parallel were laid down in a sea which transgressed over a rugged topography in which many kinds of rocks were exposed.

Early Tertiary Phase. The Paleocene and Eocene periods were marked by an important subsidence during which the sea, with some protruding insular areas, covered the western flanks of the peninsula. . . . It appears that the sea may have first occupied the coastal regions of the northern part of the peninsula from about 31° 30′ N. Lat. southward to the 27th parallel. . . . The southern margin of the sea in Baja California at that time appears to have been near the isthmus of La Paz, and the sea may have extended across the isthmus to the present lower gulf. . . . [See Fig. 30.5.]

The back country must have been of moderate elevation, well watered, thus supporting large streams, and the climate was tropical as indicated by the faunas.

An emergence near the close of Eocene time marks the beginning of a period of erosion and local folding of the Tepetate formation. The contact between Eocene beds and the overlying Miocene appears to be almost conformable where observed near the axis of the Baja California syncline, but in the Western Cape region the unconformity between the Tepetate and Miocene is more important, indicating that the earlier movement along the western marginal uplift continued in the post-Eocene.

If the sandstones at Santa Gertrudis east of Desierto de Santa Clara prove to be Tepetate in age, they probably represent the eastern limit of the formation in that area. They are overlain directly by Upper Miocene volcanics, thus indicating post-Eocene pre-Lower Miocene uplift along the axis of the peninsula near the eastern marginal uplift.

After the deposition of the Tepetate and before Miocene time, volcanism of some importance must have broken out, for the granitic rocks underlying the Ysidro beds in the Southern Cape region are intruded and in places covered by volcanic rocks; furthermore, east of San Ignacio Lagoon the basal light conglomerate of the Ysidro formation, resting on the Tepetate with slight unconformity, contains pebbles of volcanic rock.

Mid-Tertiary Phase. The depressed area (in early and middle Miocene) probably covered the synclinal region from a point in the desert area northwest of Purisima and southeastward to the Isthmus of La Paz; it probably was bordered on the west by the uplifted granitic areas at Bahia de Magdelena, which protruded as islands in the sea. The eastern extension of the marine invasion may have occupied the east coast of the peninsula from Punta San Marcial to La Paz and extended well into the adjacent gulf. . . . This possible eastward marine transgression, insofar as known, is the first Tertiary sea to have occupied any part of the gulf coast, except for the period during which the *Cornwallius* beds were deposited. [See Fig. 30.5.]

The upper Ysidro submergence in the southern area appears to have been a continuation of that which allowed the deposition of the lower shale member. It was important and widespread—much more so than the preceding. The

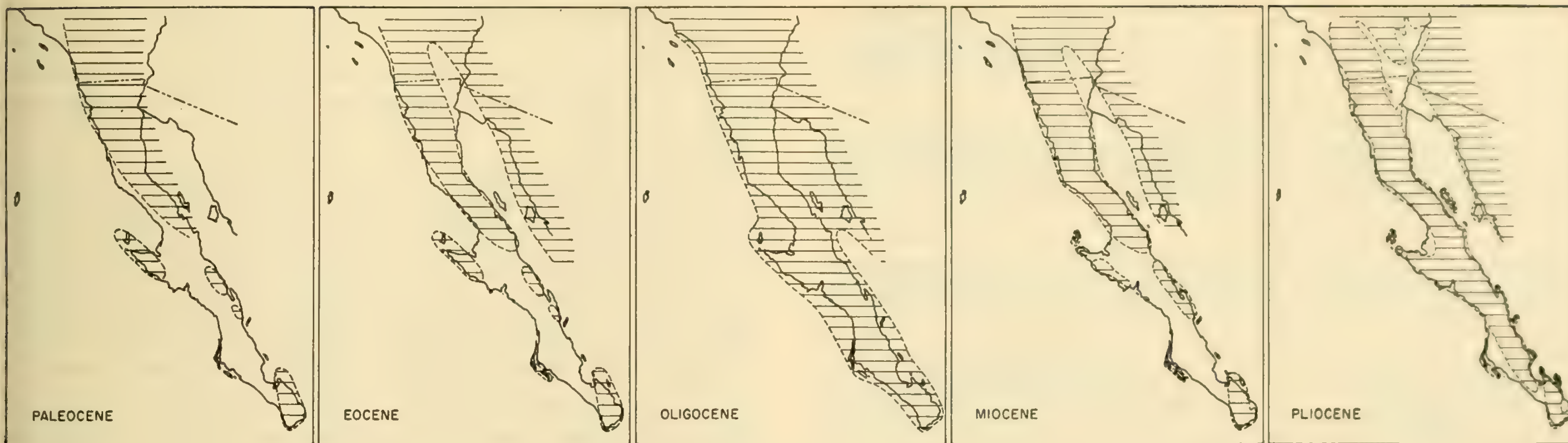


Fig. 30.5. Paleogrography of Baja California during the Tertiary, after Durham and Allison, 1960. Ruled areas denote land. The Oligocene beds of Beal are earliest Miocene on the basis of the megafauna, according to Durham and Allison.

Purisima region was again submerged. The San Ignacio area and probably much of Desierto de Santa Clara, much of the Western Cape region, and part of Isla Cedros suffered their first Miocene submergence. The northern limit of this sea may have been some place north of the 28th parallel. The eastern limit of the sea extended along the west side of the sierras, beginning not far west of Las Tres Virgenes, and crossed the peninsula to the gulf coast near Punta San Marcial.

Some structural considerations indicate that much of the lower gulf was occupied by the Ysidro sea. For example, the Southern Cape region probably was an elevated block from early Cretaceous, as it appears that in Eocene, and probably in Cretaceous time, the Isthmus of La Paz marked the southernmost limit of marine invasion, and no sedimentaries are known to have been deposited on it until Ysidro time.

The La Paz fault is deeply significant from the standpoint of the geologic history of the gulf and of the peninsula. Downthrow on the west side allowed the deposition of Tertiary and perhaps Cretaceous beds from the Isthmus of La Paz northwestward, and the northerly extension of the fault may have been a factor in severing the peninsular structural block from the old land mass. . . .

After the deposition of the Ysidro formation the peninsular area was elevated; its western margin may have been roughly coincident with the western marginal uplift and the eastern side bounded by the ancestral gulf over a part of its length, but parts of the near-shore insular area, from about 27° 30′ N. Lat. to La Paz west of the Ceralbo fault zone, were still a part of the land area. The northern half, which stood above water during Ysidro time, was further elevated, and areas of Ysidro sediments were elevated sufficiently to allow considerable erosion, especially along the margins of the peninsular area. The synclinal and some other areas appear to have suffered but minor erosion, as at many places there is little evidence of unconformity between the Ysidro and the overlying Comondú rocks.

Late Tertiary Phase. Volcanism broke out in late Miocene time and, in places, has continued down to the present. The Comondú formation of the peninsula is thick and made up of many kinds of rocks of volcanic origin. The northern half of the pensinsula must have been out of water, but the southern half was largely a site of deposition. The volcanism is probably related to the Báucarit sedimentation and volcanism of Sonora. See second from top section in Fig. 30.6.

The Baja California syncline was gently depressed; the Isthmus of La Paz elevated; the Western Cape region became a part of the peninsula, if formerly separated from it; the northern half of the peninsula had not reached its present

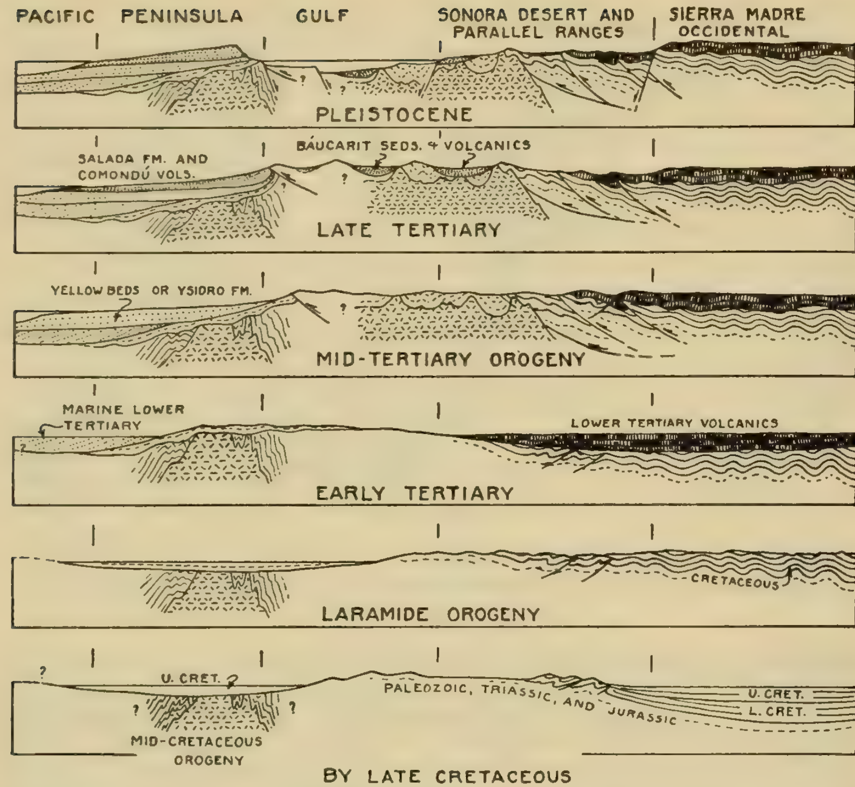


Fig. 30.6. Evolution of the provinces of western Mexico from late Cretaceous time to the present. Diagrams are highly idealized.

height; faulting and folding probably occurred along the east side of the peninsula; and further movement may have occurred along the Ceralbo and Bellenas fault zones and along unmapped faults in the adjacent sea bottom.

Some of the islands probably continued as independent structural units, the upward or downward movement of one not necessarily being coincident with or dependent upon the movement of another or with that of the peninsula as a whole. This is indicated by steeper dips in the Comondú on many of the islands than on adjacent parts of the peninsula, though some or much of this deformation may have occurred in the Pleistocene. The Comondú often shows more deformation on the east coast than farther west, which suggests that the major structural forces were more effective on that side. If true, that may have resulted in the first westward tilting of the peninsula.

The Pliocene history is complex and not well understood (Beal, 1948). Pliocene deposits in places indicate sea-level deposition, but since now observed at elevations over 1000 feet, late Pliocene or Pleistocene uplift must be postulated. In places the uplift is believed to be post-Comondú but pre-Pliocene.

An area of greater significance is the known marine Pliocene at Santa Rosalia, which has been elevated to about 500 feet, possibly more. According to information from Mr. Ivan F. Wilson, the underlying Comondú formation occurs in a series of fault blocks cut by southwesterly dipping faults. As these faults are probably of pre-Pliocene age, the sierra southwest, composed principally of Comondú rocks and rising to more than 5000 feet, must have reached nearly that elevation at the time of the post-Comondú uplift and deformation, but in pre-Pliocene time. It is doubtful if any major change in the relative elevations of the Comondú and Salada areas has been caused by erosion and deformation.

In the diagrams of Fig. 30.6, the post-Salada and post-Báucarit disastrophism is indicated as due to compression, and the Gulf of California had not yet come into existence. According to Beal, however, some faulting had probably occurred in mid-Tertiary time, and not all of the downfaulting of the gulf and the uplift of the peninsula took place in the Pleistocene, as illustrated. It is certain, however, that a great deal of the displacement that shaped these major elements is post-Salada.

Quaternary Phase. The submarine canyons on the continental margin have been regarded as of subaerial erosion, and hence to represent a great emergence, according to Beal, in postfaulting time. This does not seem necessary, however, because when once deeply submerged, the form remains little changed, and the canyons may be of considerable antiquity. Better understood is evidence of a great Pleistocene submergence. Terraces and marine shells lead Beal to conclude that

. . . there seems little doubt that the sea level rose at least 1600 feet, and Wittich (1920) believed it rose to about 3000 feet. If the depression of 1600 feet was uniform throughout the full length of Baja California, the peninsula would have been only about two-thirds its present length with a string of islands extending southeastward.

Johnson (1924), in his study of the fauna and flora of Baja California, states "For some reason the fauna and flora were subjected to a crisis during Pleistocene, and all but a few vertebrates were destroyed." This wholesale destruction might have resulted from the submergence of the peninsula, indicated by the presence of sea shells at considerable heights.

The following emergence of equal magnitude may still be going on in places.

Volcanoes were active during the Pleistocene and have continued their activity to Recent time. Isla Tortuga is the youngest island in the gulf; it erupted from the gulf floor about 6000 feet deep and reached an elevation above the gulf of more than 1000 feet. Its poorly eroded surface and lack of vegetation vouch for its youth. Las Tres Virgenes are said to have been active in historic time. On the west slope of the sierras there are many Quaternary craters and cones, and at San Quintin the volcanic flows, according to Woodford (1928), may be in part historic. Cerro Prieto, near Volcano Lake (Mexicali quadrangle), is a small perfect crater probably formed in Recent time.

Movement along older fault lines continued during the Quaternary, and probably many new crustal breaks were initiated, whether in early or late Pleistocene is not known, but one may assume that much of this activity occurred at the time of the Middle Pleistocene revolution of California. Movement still continues along some of the fault zones in both California and Baja California. The existence of zones of faulting which border the peninsula is indicated most strongly by recent phenomena, though activity along some of these zones probably has been nearly continuous from some remote time.

Quaternary uplift has increased the height of the mountains of the peninsula, rejuvenated streams in regions of low relief, and exposed a considerable area of partially consolidated beach material to erosion, with the resulting development of a coastal slope which appears from a distance to be a plane surface, but which is really an intricate pattern of small arroyos and narrow ridges. Recent erosion has cut deep canyons into the rocks of the peninsula and reduced the height of its mountains, while alluvial deposition has in places half buried some of the ranges in fans of detritus derived from them. The wind has assisted in sculpturing some of the softer rocks in regions of rugged topography, and the outlines of the topography are softened by the addition of aeolian material in the broad low desert regions; giant sand dunes, or médanos, are numerous and cover large areas in the desert regions.

At the head of the Gulf of California the Colorado River formed an enormous delta over which it flowed alternately into the gulf and then northward into the Salton Sea, making what is now the Salton Basin into a fresh-water lake. The Coahuila Indians have handed down legends about this diversion.

GULF OF CALIFORNIA

Shepard and Emery (1941) and Beal (1948) consider the Gulf of California to be a downfaulted trough complementary to the uplift of the peninsula of Baja California. King (1939) has suggested a relation of the faults of adjacent Sonora to those of the Gulf. Beal has described the submerged topography as follows:

> The northern quarter of the gulf is shallow—at no place more than 600 feet deep. The deepest parts of the gulf south of the 30th parallel appear to lie west of its center, and thus probably before the floor was deformed by so much faulting it simulated, in some respects, the westward-tilted block of the peninsula, suggesting an extension of the basin and range structure of the Sonora area.
>
> The east side of the gulf appears not to have been affected by faulting; the gulf floor slopes gently from the Sonora coast to the irregular escarpments near the center of the gulf. The most important of these is the great submarine cliff nearly 6000 feet high between the 25th and 26th parallels. Between the 24th and 27th parallels there are many irregularities in the submarine topography between the Ceralbo fault zone and the east coast of the peninsula, but most of them lie west of the Ceralbo fault zone.
>
> Between 30 and 40 islands varying in size from Isla Angel de la Guarda, between 75 and 80 kilometers long, to very small ones, rise above the surface of the gulf, some to surprising elevations. Other islands such as Consag Rock, San Pedro Mártir, Ceralbo, and Santa Catalina have the appearance of wedges uplifted from the granitic floor of the gulf or as stocks or spurs still attached to the granitic batholith.
>
> Much of the south half of the gulf is occupied by a remarkable depression in the sea floor, extending 400 kilometers southeastward from a point east of Isla Tortuga. It widens into enormous proportions at places and becomes narrow in others, with the closing depression contour 5400 feet below sea level. This great depression area is occupied by three separate smaller basins, the largest and deepest (10,740 feet) of which lies in the center of the gulf between 25° and 26° N. Lat.
>
> A distinctive depression about 250 kilometers long, the origin of which can reasonably be assigned only to faulting, separates the Angel de la Guarda group of islands from the peninsula. The deepest part of the trough is about 5100 feet and lies adjacent to Isla Sal si Peudas. The closing depression contour is 1200 feet below sea level, thus furnishing a long narrow basin, nearly 4000 feet deep, which widens at its north end. A line indicating the east boundary of the graben is called the Ballenas fault zone, the northwestward extension of which may lie farther west than shown and join the northwestern extension of the Ceralbo fault zone. [This fault, or fault zone, has been drawn on Fig. 30.6 as the western boundary of the depression which the writer interprets as a graben.]
>
> From the configuration of the gulf floor, there seems good evidence of a fault east of the Isla Ceralbo [Fig. 30.6]. At the sea bottom, north and east of this island, is a submerged island nearly three times as long as Ceralbo, with its crest approximately 1000 feet below sea level and rising about 2500 feet above its base. Topographically, the submerged island appears to have been once a part of Isla Ceralbo, and both apparently a part of the Southern Cape

region, but the submerged island is now separated from Ceralbo by a deep, narrow basin with its bottom 6000 feet below sea level and a sill depth of 4800 feet. Immediately northwest of Isla Ceralbo there is a smaller submerged hill about the same size as Ceralbo; its crest lies only 600 feet below sea level, and it may originally have been a part of the same mountain mass. If facts can finally be collected on the structure of the gulf floor east and northeast of the Southern Cape region, they will probably show that the deep basin immediately east of San Jose del Cabo has been caused by north-south faulting parallel to the La Paz fault, and that the deep narrow basins farther north owe their origin to northwest-southeast faulting, with the same structural trend as the gulf trough.

In seeking for the cause of the broad, deep basins in the gulf below the 28th parallel, one may conjecture that they are probably structurally depressed, wedge-shaped blocks, bounded by faults. Ballenas Channel and the depression east of Isla Ceralbo, both of which appear to be grabens, may have originated in the same way. If they were deep troughs with open ends, instead of elongated steep-sided basins, their unusual depths might have been attained by the erosive action of the gulf currents. It appears that their great depth as basins, however, can logically be explained only by assuming the basins to be the apices of structurally depressed wedges.

Tertiary sediments in the Salton basin and farther northwest may be very thick, and Beal suggests that basement rocks under the north end of the gulf trough may be 25,000 to 30,000 feet below sea level. It is generally recognized that the Colorado delta has contributed much toward filling the trough and making the present floor shallower.

SIERRA MADRE OCCIDENTAL

Early Tertiary Phase

According to R. E. King (1939), the Sierra Madre Occidental takes form south of the international boundary by the coalescing of mountain ranges which, in southern New Mexico and Arizona, are more or less isolated. South of the boundary, the plains between the mountains become narrower, and the volcanic rocks spread out in a broad plateau. The western edge of the plateau, at an elevation of 6000 feet or more, breaks off toward the Gulf of California in lofty escarpments which are trenched by most impressive gorges. West of the Sierra Madre proper, high ranges are separated by long, narrow valleys. Still farther west, bordering the gulf, low mountains are separated by broad plains, as in the Basin and Range province of southwestern United States.

The three geomorphic divisions have been called, by King, the Sierra Madre Occidental province, the province of parallel ranges and valleys, and the Sonoran Desert province.

The Sierra Madre Occidental has generally been assumed to be a structurally simple plateau of flat-lying lavas overlying a basement of sedimentary rocks and ancient granites, but a reconnaissance survey by R. E. King (1939) has added greatly to our knowledge of the region and revealed a complex structural history. The rocks studied by King have been much folded and faulted and are intruded by numerous plutons of fairly large size. There are several periods of deformation, but only those of the Tertiary can be deciphered with any assurance. Two unconformities in the Tertiary mark times of important mountain building. The structural features produced by the Tertiary episodes of deformation trend in general north-northwest, and produce a conspicuous alignment of rock outcrops and ridges.

The effects of the Laramide revolution have already been mentioned in connection with the Mexican geosyncline. See lower two sections of Fig. 30.6. Early Tertiary volcanic rocks spread out over much of the surface of western Sonora but reached their greatest development in the plateau section of the Sierra Madre Occidental. See third section from bottom of Fig. 30.6. In the plateau section, the underlying Mesozoic rocks are probably greatly deformed, for such disturbance is evident along the western edge of the plateau and in the few inliers within the plateau and in the Nevadan type rocks of Baja California. The later or post-volcanic deformations strongly expressed to the west in the parallel ranges and Sonoran Desert have, however, scarcely affected this region. Over wide areas, the volcanic rocks are more than 5000 feet thick and are flat or gently tilted. They consist of flows and pyroclastics with basalts dominant in northern Sonora (Imlay, 1939) and more acidic types most voluminous in central Sonora (King, 1939). The volcanic layers were then uplifted epeirogenically thousands of feet, evidently, because an erosion surface developed to maturity on them. It is now deeply dissected by the present cycle of erosion (King, 1939). See fourth section from bottom, Fig. 30.6.

Toward the west the plateau gradually loses its structural simplicity.

Within the barranca section (great gorges indenting the west-facing escarpments) not only are the plateau summits largely destroyed by erosion, but the volcanic rocks are also broken by faults that belong to later deformational phases.

Mid-Tertiary Phase

Parallel Ranges and Valleys. After the early Tertiary eruptions, there was a vigorous phase of mountain making that is known principally in the province of parallel ranges and valleys and in the Sonoran Desert. Within the plateau section of the Sierra Madre, the volcanic rocks were only gently folded, and over wide areas they still remain nearly horizontal. This gentle folding contrasts with the strong disturbance of the Cretaceous and other Mesozoic rocks, where they can be observed beneath, and indicates that the Laramide orogeny was greater than the mid-Tertiary in the Sierra Madre proper.

Farther west, as in the province of parallel ranges, folds and thrust faults occur that can be assigned to the mid-Tertiary deformation, which here exceeds the Laramide. The mountains probably began to assume their present form at this time.

The ranges are generally bordered by faults. North of the 28th parallel, the faults are high- and low-angle thrusts. To the south, steep normal faults predominate. They are not all, however, of the mid-Tertiary disturbance; some are late Tertiary.

Accompanying the mid-Tertiary orogeny were vast intrusions of granite and other plutonic rocks, which ascended through the Paleozoic and Mesozoic rocks and, in places, penetrated the early Tertiary volcanics. Some of the faulting started at this time, because several of the thrust faults that break the early Tertiary volcanic rocks of the barranca section are cut off by granite intrusions. See fourth cross section from bottom of Fig. 30.6.

Sonoran Desert Province. North of latitude 28° 30′ N., a large proportion of the detached mountain ranges in the Sonoran desert province consists of Paleozoic and Mesozoic sedimentary rocks. To the south in the unmapped area that extends to the coast, they consist of volcanic rocks and granite. The ranges of sedimentary and volcanic rocks appear to be only detached roof pendants in a vast granite batholith or group of coalescing batholiths (King, 1939). They probably represent the lowest part of the roof at the end of the period of intrusion. Nearly all the nongranitic rock in the ranges is cut by apophyses of granite.

The granite intrusions have complicated the pregranite structure of the sedimentary and volcanic rocks by metamorphosing and shattering them close to the contact and by jointing them excessively for some distance from the contact. Alternations of competent and incompetent strata, such as are found in parts of the Paleozoic and the Jurassic Barranca formation, shows such a confusion of dips and small faults that it is very difficult to work out the main structural features. Only the most massive, resistant formations, such as the Permian limestone and the upper part of the Barranca formation, show the structure clearly; and even these only at some distance from the nearest granite contact (King, 1939). Despite these confusing relations, King finds some of the larger features of the structure plain. The mountains in part are clearly upfaulted. Some still preserve the form of tilted fault blocks, although considerably modified by erosion. Some of the depressions are downfaulted, and some overthrusting is present in the Sonoran Desert.

Baja California. The bulk of the Tertiary formations in Baja California are the result of orogeny to the east in Sonora. This is particularly true of the "yellow beds" (Darton terminology). They are present in great volume and coarsen eastward. From the relations that Darton depicts, the yellow beds are the great orogenic deposit in the southern half of Baja California, and if they are the late lower and middle Miocene Ysidro formation of Beal, then the mid-Tertiary orogeny of King in Sonora is probably dated by them.

In Sonora itself, the next youngest formation after the disturbance is the Báucarit of late Pliocene or Quaternary age. It occupies the depressions between ranges and probably was deposited some time after the orogeny. See second section from the top of Fig. 30.6. The yellow beds were upturned in places, eroded, and then covered with sands, conglomerates, agglomerates, and basalt flows. Since these capping deposits are late Miocene (?) and Pliocene in age (Beal, 1948), it appears that the yellow beds were deposited in a hurry and then immediately somewhat de-

formed. Both the orogeny that resulted in their deposition and the impulse that deformed them might, therefore, be considered parts of the same phase until more information is available.

Late Tertiary or Early Pleistocene Phase

After the mid-Tertiary orogeny in the province of parallel ranges and valleys and in the Sonoran Desert, the Báucarit formation was laid down in the structural depressions between the uplifted mountain ranges. A moderate recurrence of volcanic activity is indicated by the basalt flows in the lower part of the formation. Interbedded with and overlying the basalts are conglomerates which were doubtless laid down as coalescing alluvial fans at the margins of the mountains. They contain fragments derived from the cores of the ranges including boulders of granite. Similar deposits overlying the yellow beds of Baja California have already been mentioned. Beal describes two formations there, the lower Comondú volcanics and the overlying Salada formation, which appear similar to the Báucarit, and correlative with it. The next structural phase postdates the Báucarit, and is of varied aspect. There was renewed volcanic activity, and the Báucarit formation was thrown into low folds and tilted.

Quoting from King (1939):

> North of the 28th parallel, the rocks of each of the high mountain ranges, from the crest of the Sierra Madre westward into central Sonora, were pushed to the west on overthrust faults which partly overrode the Báucarit formation, lying in the valleys next to the west. Some minor faults were thrust to the east. The strong thrusting and the gentle warping of this orogenic epoch suggest that the strata of the mountains had already become so consolidated by previous folding and igneous intrusions that they could no longer yield to lateral pressure by folding. The greater amount of thrusting north of the 28th parallel may be due to the greater thickness of Paleozoic and Mesozoic sedimentary rocks in that region.
>
> The normal faults extensively developed south of the 28th parallel and farther west in central Sonora were somehow related to the thrust faults. At La Colorada these offset the plane of an overthrust fault, but both here and to the south they have the same north-northwest trends as the overthrusts and thus may have taken their form from the same forces. In the province of parallel ranges and valleys, the localization of overthrusts north of the 28th parallel and of normal faults to the south of it suggests that the orogenic forces, although dominantly compressional, produced local areas of tension.
>
> During rather recent geologic time, a mature erosion surface of low relief was developed in the lava country along the crest of the Sierra Madre. After its formation, the area was greatly uplifted, and streams draining to the west deeply intrenched their courses, forming the tremendous barrancas of the western flank of the Sierra Madre. It is not entirely certain when this uplift took place, but the great height of the surface above low country not far to the west strongly suggests that it was raised by faulting on the west side of the plateau. This faulting may have been the post-Báucarit thrust faulting, or it may have been a renewed movement at a later time along the same trends.

King cannot date the elevation of the lavas of the Sierra Madre Occidental with accuracy, but he believes the elevation was due to faulting in post-Báucarit time. It seems probable, therefore, that the elevation of the Sieras occurred at the same time as the sagging of the Gulf, and that they are parts of the same fault block system. The differential movement, as estimated from the bottom of the Gulf to the crest of the Sierras, is about 12,000 feet.

The position of Baja California in the regional tectonic plan is treated in Chapters 29 and 31.

31.

MIDDLE AND LATE CENOZOIC SYSTEMS OF THE CENTRAL CORDILLERA

GENERAL DIVISIONS AND THEIR CHARACTERISTICS

For structural purposes it seems best to treat the middle and late Tertiary mountain systems in the central part of the great Cordillera of North America in three divisions, namely, the Basin and Range system of southern Oregon, eastern California, Nevada, western Utah, and northwestern Arizona; the Sonoran–Chihuahua system of desert ranges in western and southern Arizona, New Mexico, and central Mexico; and the system of great trenches in central Utah, eastern Idaho, western Wyoming, western Montana, and British Columbia. The first two divisions are generally included by the physiographers in the Basin and Range province, and the third has generally not been distinguished from the Laramide Rockies on whose folds and thrusts its fault-made trenches are superposed.

The Basin and Range system is one generally of north-south-trending basins and ranges, with the majority of the ranges probably blocked out by high-angle faults. The distinctive features of the province, according to Fenneman (1931), are "isolated, nearly parallel mountain ranges (commonly fault blocks) and intervening plains made in the main of subaerial deposits of waste from the mountains. These deposits, although locally absent, are often very deep and are generally unconsolidated."

The boundaries of the Basin and Range system are shown on the map of Fig. 31.1. The Great Basin of internal drainage, the Mojave Desert, and the Salton trough are the chief regions here included. The Basin and Range province is bounded on the west by the Sierra Nevada, on the east by the Wasatch Mountains and High Plateaus of Utah, and on the north by the Malheur plateau and Snake River lava plains. The narrowing south end has been arbitrarily defined by Nolan (1943) to have the San Andreas fault on the west and the Colorado River on the east. The physiographic section of the Great Basin, called the Sonoran Desert by Fenneman, includes large areas on both sides of the Colorado River; and the Basin and Range system is probably continuous across it to the desert ranges of southern Arizona.

The desert ranges of Arizona, New Mexico, and part of the Mesa Central of northern interior Mexico are somewhat similar to those of the Great Basin in being rudely parallel and separated by basins filled in part or completely by alluvium. Those in southern Arizona stretch northwestward across the southern and southwestern part of the state. They converge haphazardly, in the southeastern corner, with basin ranges of New Mexico which extend northward through the central part of the state. Together, the ranges of block-fault character of Arizona and New Mexico extend southward into Mexico through the state of Chihuahua to Durango, Coahuila, Zacatecas, and San Luis Potosi. The sources of information on this great region are a few detailed studies of wide separated areas. Large parts of it have never been reported on, so the concept is not secure that it is everywhere a mountain system whose present

form is due largely to middle and late Tertiary deformation. Literature on the structure of the desert ranges of the western part of the Plateau Central of Mexico is almost nonexistent.

All the great trenches of central Utah, eastern Idaho, western Wyoming and Montana, and British Columbia are probably fault valleys and of middle and late Tertiary age. They extend as a narrow belt from the High Plateaus of northern Arizona and central Utah through the Wasatch Mountains in Utah and northward along the boundary of Idaho and Wyoming to the Teton Range and Jackson Hole in northwestern Wyoming, thence northwestward as a wider belt through western Montana to the great trenches of British Columbia. See the map, Fig. 31.1 for boundaries. Much also remains here to be worked out; but sufficient is known, it is believed, to compose these great valleys into a structural system and to treat them collectively as such.

With few exceptions, the middle and late Tertiary high-angle faults of the Great Basin and the folds and thrusts of southern California are superposed on earlier Nevadan or Laramide structures.

BASIN AND RANGE SYSTEM

Evidence of Faulting

Four types of evidence have been used to prove that the individual ranges in the Great Basin are bordered by block faults: physiographic evidence, stratigraphic evidence, exposure of a fault plane, and presence of recent fault scarps along the range fronts. As the boundary between mountain and valley blocks is commonly concealed by the alluvium accumulating in one or more closed basins, the second and third types of evidence are rarely found; for most places physiographic evidence has been called upon to determine the existence of a fault block.

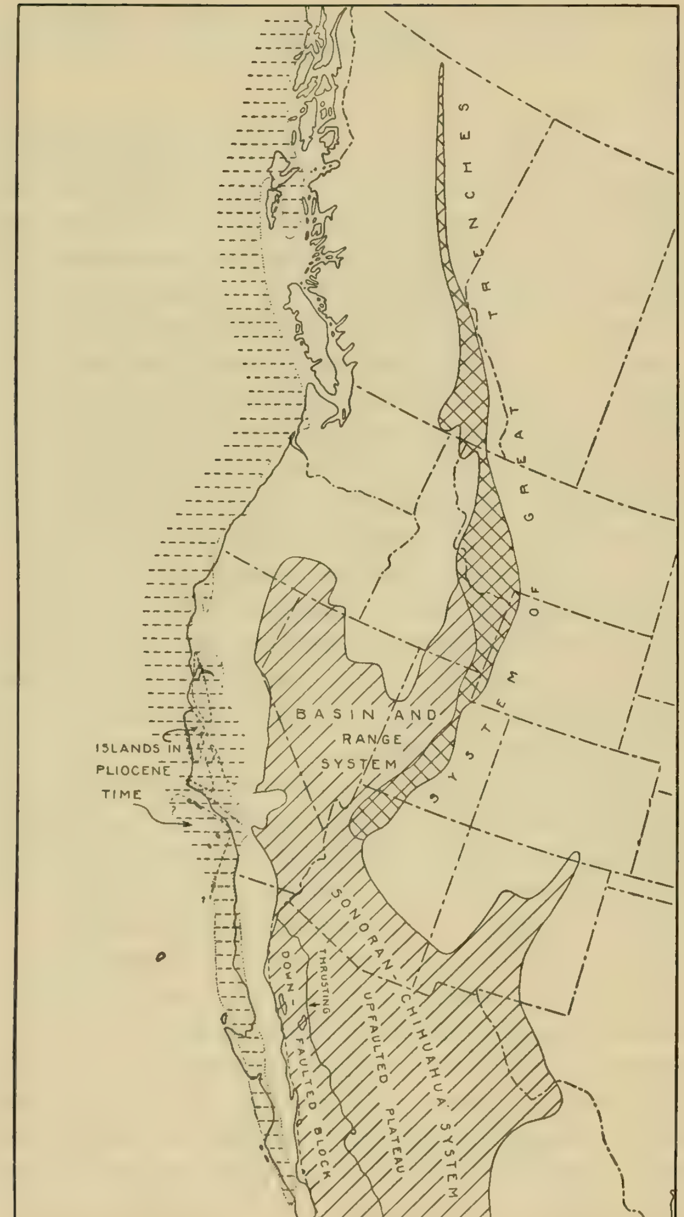


Fig. 31.1 Tectonic map of the western Cordillera in late Miocene, Pliocene, and early Pleistocene time. The sediments along the Pacific are late Miocene and Pliocene in age. They are horizontally dashed. The obliquely ruled area denotes the Basin and Range and Sonoran-Chihuahua systems, the faulting of which took place chiefly in Pliocene and early Pleistocene time, although in places it started earlier and lasted longer, even to the present. The cross-ruled belt is the system of great trenches. Miocene and Pliocene basin deposits are common in all three fault systems.

The following kinds of physiographic evidence have been used: the front of the range is linear and cuts indiscriminately across the rock structure; the range rises abruptly from a waste-filled valley; many steep, narrow V-shaped ravines cleave the mountain block and open abruptly onto the gravel fans of the valleys, and triangular facets are aligned along the mountain front. Major valleys cutting through the ranges are generally absent; mature topography or thin caps of volcanic rocks mark summits or back slopes of the ranges; landslides are common along the range fronts; hanging valleys are present on some range fronts; and the lowest point in the adjoining valley is close to the scarp along the range front.

Blackwelder (1928) has reviewed these and other proposed criteria and has pointed out that several of them are equally applicable to exhumed or "fault-line" scarps. He regards the following features as positive evidence of true fault scarps: (1) lack of correlation between rock resistance and surface form; (2) rift features; (3) alluvial deposits on the downthrown block thickest near the fault line; (4) lake or sink close to the scarp base; (5) alluvial fans abnormally small; (6) frequent severe earthquakes; (7) displacement of an older topographic surface; (8) dislocation of Recent or late Pleistocene formations; (9) basal scarplets; (10) warped terraces in the canyons; and (11) the fault plane identified as forming part of the scarp face. Nolan (1943) comments that some of these features are of relatively little value because of their infrequent occurrence (item 10, for example) or because of the absence of adequate information (item 3); and others, such as item 6, are of questionable dependability. Other observers would probably regard additional features as equally valid evidence.

When critically used there is little doubt that physiographic evidence alone is adequate and diagnostic. In many places, however, use of evidence of this type has resulted in a failure to distinguish between fault scarps and fault-line scarps; and there has even been a tendency to consider that any elevated block with an approximate linear trend is necessarily a fault block.

Stratigraphic evidence of faulting along the borders of ranges is generally difficult to find because valley fill commonly conceals the downthrown block. Stratigraphic proof of faulting has been found in the Humboldt Lake and adjoining ranges, Nevada (Louderback, 1904); the Lake Tahoe region, California–Nevada (Reid, 1911); the Oquirrh Range, Utah (Gilluly, 1928b); the Warner Range, California (Russell, 1928); the Wasatch Range, Utah (Eardley, 1934); the Deep Creek Range, Utah (Nolan, 1935); the Boulder Dam region, Nevada (Longwell, 1936); and the Comstock Lode, Nevada (Gianella, 1936). In other places faulting along the range front has been inferred from the presence of parallel step faults within the range (Fuller and Waters, 1929).

In a few places, no evidence of faulting at the contact between the rocks that form the ridges and the Tertiary sedimentary beds that underlie the valleys is apparent. Ferguson and Cathcart (1924), however, have interpreted similar occurrences in central Nevada as the result of sedimentation on the downthrown block, which overlapped the outcrop of the fault.

Actual exposures of faults bordering the ranges have been made accessible by artificial excavations, but in a few places they have been revealed by erosion. The Wasatch fault has been located by Pack (1926) and Eardley (1934), several faults along the west edge of the Oquirrh Range have been located by Gilluly (1928b), several Pliocene faults in southern Nevada have been located by Longwell (1936), and additional faults in central Nevada have been located by Ferguson. In the region studied by Longwell a considerable vertical extent of the fault was revealed, and here at least the dip of the fault steepened upward; at the other localities fairly steep valleyward dips prevail, ranging from 50 to 72 degrees.

Small scarps formed by recent faulting, called piedmont scarps by Gilbert (1928) or fan scarps by Longwell (1930), correlate closely with the scarps bordering many of the basin ranges. This was first pointed out by Russell (1884), and since that time these recent scarps have been commonly considered to indicate the presence of persistent faults. Many of them have been recognized throughout the Great Basin, those in the Lahontan and Bonneville basins by Russell (1885) and Gilbert (1890, 1928a); those along the Sierra Nevada by Hobbs (1910), Lawson (1912), and Knopf (1918); those in central Nevada by Jones (1915), Page

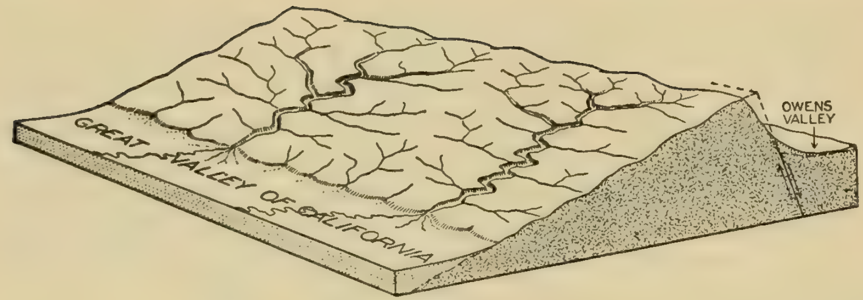


Fig. 31.2. Generalized diagram of part of tilted Sierra block. The great fault fractures that separate the Sierra block from the Owens Valley block, on the east, are shown by a single line. The height and slant of the Sierra block are much exaggerated. The streams are shown in their characteristic arrangement, the main rivers flowing down the western slope but many of their tributaries in directions approximately at right angles to them. No specific streams are represented. In front is a strip of the Great Valley of California, whose thick layers of sand and silt, derived from the elevated part of the Sierra block, bury the sunken part. At the back is a strip of Owens Valley, veneered with a thinner layer of sediment. After Matthes, 1930.

(1935), Gianella and Callaghan (1934); those in southern Nevada by Longwell (1930); and those in southern Oregon and northeastern California by I. C. Russell (1884) and R. J. Russell (1928). In some places these scarps have clearly been developed between the hard rocks of the range and the gravel of the valley. Commonly, however, they are found in the gravel some distance from the range front, and tend to be more irregular than the front in plan. Although most of the recent scarps lie at or close to range fronts, some are also found in the intervening valleys (Gianella, 1934; Gianella and Callaghan, 1934) and within the mountain ranges (Callaghan and Gianella, 1935). Many of them are accompanied by hot springs (I. C. Russell, 1884) or are coincident with volcanic cones (Knopf, 1918).

Nature of Block Faults

The Sierra Nevada Range is a westward-tilted fault block. See Fig. 31.2. The faults that border the east front of the range are staggered in map plan. Along the great escarpment that faces Owens Valley there may be a single fault, or perhaps a set of closely spaced parallel faults; but farther north the successive offsets in the front of the range indicate the existence

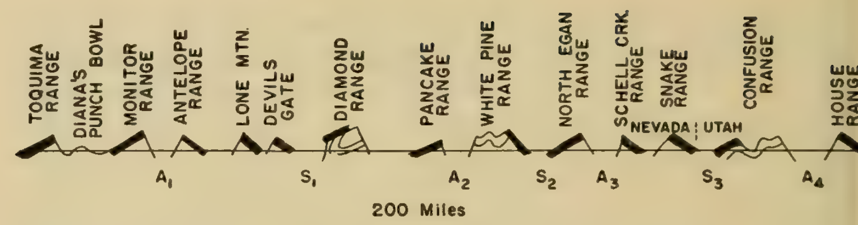


Fig. 31.3. Diagrammatic section from central Nevada to western Utah. Reproduced from Osmond, 1960. Solid black represents tilted Tertiary volcanic deposits. The ranges can be interpreted as being the remnants of four large anticlines, A, and intervening synclines, B.

of discontinuous northward-trending fractures that replace one another at intervals, thereby splintering the northwestward-trending margin of the block on a large scale. From the neighborhood of Lake Tahoe, which itself lies in a trough produced by the subsidence of a great splinter, long lines of faulting diverge in northerly directions, each marked by an escarpment of its own. Northward the eastern margin of the Sierra block becomes progressively more irregular, and the displacements are distributed over a belt that broadens gradually to a maximum of 50 miles. Some of the escarpments measure but a few hundred feet in height and the highest do not exceed 2000 feet (Matthes, 1930).

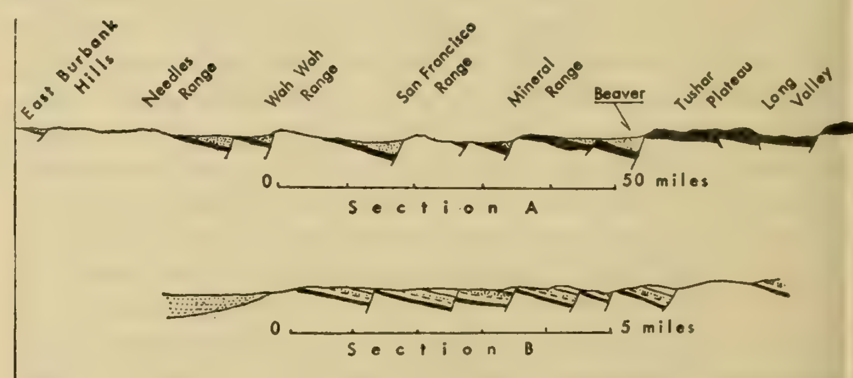


Fig. 31.4. Diagrammatic sections of southwestern Utah. Section A from the Nevadan boundary to the High Plateaus, and section B from the Escalante Desert to the High Plateaus. The black areas represents Tertiary volcanic deposits. Reproduced from Mackin, 1960.

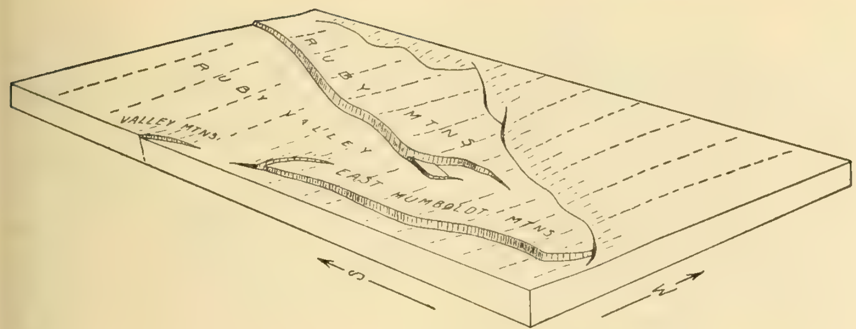


Fig. 31.5. Block diagram showing nature of crustal deformation by block faulting in the Ruby–East Humboldt Range, Nevada. The view is to the southwest. The block is about 50 miles long (S) and 30 miles wide. (W). The diagram is approximately to scale with the maximum throw of the faults about 6000 feet (Sharp, 1939). The faults acquired their present displacement by four stages of movement from upper Miocene to the Pleistocene.

Studies by Hudson (1955) indicate that the uplift of the Sierra Nevada is not due to simple tilting of a rigid block. An important zone of faulting about midway between summit and the western edge of the range divides the range into two blocks of deformation, and Hudson, from gradient calculations, thinks there may be five separate bocks of adjustment.

Most of the blocks throughout the Great Basin are rotated or tilted. Study Osmond's representation in Fig. 31.3, which includes thirteen uptilted blocks and probably a number of down-dropped additional blocks in a distance of 200 miles. In the eastern part of the Great Basin Mackin, working with the ignimbrite sheets, shows a series of blocks all rotated in the same direction. See Fig. 31.4. Some blocks, however, are horstlike, such as the Ruby Mountains and East Humboldt Mountains (Fig. 31.5).

Mapping in the Wasatch Range (Eardley, 1939, 1944) indicates that a master fault, the Wasatch fault, 115 miles long with displacement of 1000 to 6000 feet, forms the eastern limit of the faulted part of the Great Basin. In places its displacement is distributed along step faults with the west side down. It is a dip-slip normal fault and dips 50 to 70 degrees west. A quasi *en echelon* pattern of smaller normal faults spreads across the thick sediments of the Pennsylvanian Oquirrh basin of west central Utah. The Basin and Range faults are not aligned with the Precambrian or Laramide structures. Neither have the crystalline rocks of the northern Utah highland, the intrusions of the Cottonwood uplift, nor the late Precambrian basins influenced perceptibly the course or the throw of the faults. The widths of the fault blocks range from 4 to 30 miles, but a fairly uniform width of 18 to 24 miles is found in the four major blocks of the area, the Wasatch, Oquirrh, Stansbury, and Cedar mountain blocks. A relief of 3000 feet or more is believed to have existed at the inception of faulting.

Age of Block Faulting

Ferguson (1926) and Ferguson and Cathcart (1924), in addition to presenting physiographic evidence that the block faulting occurred at different times, found that similar faults, though without present topographic expression, both preceded and followed the deposition of sediments belonging to the Esmeralda formation (late Miocene and early Pliocene). The conclusion that these earlier faults were of the same character as the later block faults is based on the fact that the Esmeralda, adjacent to the pre-Esmeralda faults, is composed of material similar to that now being deposited in the fans along range-front scarps, and further that at least some of the topographically expressed faults have followed the lines of these early faults. Westgate and Knopf (1932) have also found evidence in the Boulder Dam region for block faulting that preceded, accompanied, and followed the deposition of his Muddy Creek formation, of questionable Pliocene age. Gianella (1936), similarly, has distinguished two major epochs of movement at the Comstock Lode.

A typical range in the north central part of the Basin and Range province for which the geology has been worked out is the Ruby-East Humboldt. According to Sharp (1939) the range consists of pre-Miocene igneous, metamorphic, and sedimentary rocks of complex structure. The adjoining basins contain deformed beds of the upper Miocene Humboldt formation. The boundary structure of the mountain block is well exposed because of dissection by the through-flowing Humboldt River. See Fig. 31.6.

This range is a westward-tilted horst, bounded by normal faults which

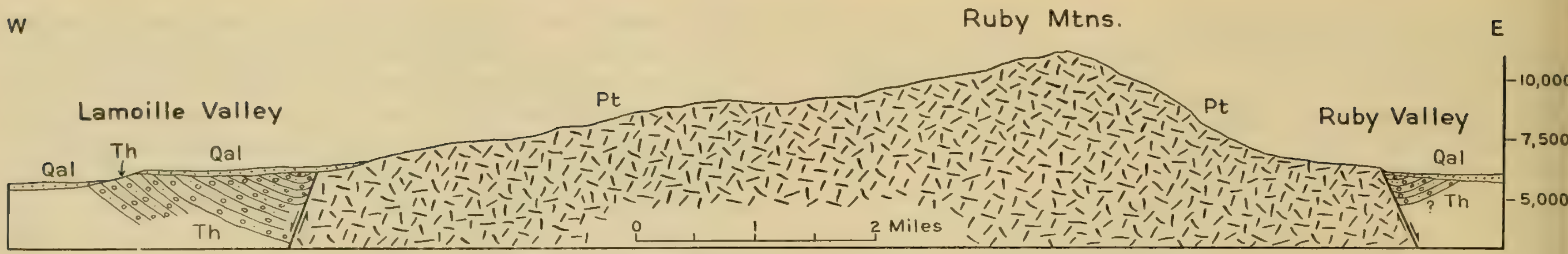


Fig. 31.6 Section across the Ruby Mountains, Nevada, showing relation of Miocene Humboldt formation to boundary faults. After Sharp (Fig. 3, 1940).

dip 60 to 70 degrees basinward. Displacements on the east boundary faults have been at least twice as great (5500 to 6000 feet) as on the west boundary faults (2000 feet). The northern termination of the mountains is due to intersection of the east and west boundary faults. The structure of the pre-Miocene rocks is discordant with the trend and shape of the range.

Five periods of basin-range faulting have been established: (1) late middle or early late Miocene, displacement small (open to question); (2) late Miocene, during deposition of the Humboldt formation, displacement larger; (3) latest Miocene to Pliocene, younger than the Humboldt formation and older than the Pliocene (?) lava, amount of displacement unknown; (4) Pliocene to Pleistocene, later than the Pliocene (?) lava and extending to middle or late Pleistocene, the period of last major uplift of the range, displacement large; (5) late Pleistocene to Recent, later than the earliest Wisconsin, displacement small.

The history of faulting in the Sierra Nevada is fairly completely known. Toward the end of the Eocene, volcanoes were intermittently active, and they emitted rhyolite lava and mud that filled the existing valleys. This volcanic activity, interspersed in an erosion cycle, continued well into the Oligocene; at the same time the Sierras were gaining elevation by vertical arching. The country lying to the east was warped and flexed; low ranges came into existence, and between them were formed wide basins in which the water collected in shallow lakes.

According to Matthes (1930), the disturbances died out at last and were followed by a long interval of relative quiet, during which most of the rhyolite and much other rock waste was stripped from the Sierra region and deposited on its western border and in the basins to the east of it. Then, presumably in the second half of the Miocene epoch, volcanic activity and earth movements began anew on a large scale. This time, the eruptions yielded mostly andesitic lava of brown, reddish, and grayish colors. Down the valleys this material flowed, sheet upon sheet, obliterating the stream beds and compelling the waters to seek new paths. In the north half of the range, the outpourings were especially frequent and voluminous; they piled up to thicknesses of a thousand feet or more, overwhelming all the features of the country save the higher peaks and crests. In the southern parts of the range, the volcanic flows were less extensive and less thick; they filled only the bottoms of certain valleys, and caused no notable displacements in the drainage system. Only the drainage basin of the Merced River, in the central part of the range, remained free from volcanic outpourings.

The crustal movements of this epoch increased the height of the Sierra region by several thousand feet and gave it the aspect of a mountain range, or rather a belt of mountains, that dominated all the country round about. Mount Lyell probably attained an altitude of about 7000 feet. Strong faulting took place along some parts of the eastern border, and the great depression in which Lake Tahoe is situated was formed by subsidence; the lava which dams the lake itself was not poured out, apparently, until after the depression was formed. The ranges and valleys of

Fig. 31.7 Wasatch fault in the north central Wasatch Range and its relation to the erosion surfaces.

the Great Basin region were accentuated, in part by warping, in part by faulting.

Next followed another long interval of repose, or relative repose, that lasted through the entire Pliocene epoch. Only feeble eruptions took place from time to time, and meanwhile the waters in the lava-covered parts of the range reorganized themselves into new rivers and cut new canyons, some of which attained depths of more than 1000 feet.

Then, at the beginning of the Quaternary period the great uplift and tilting commenced that gave the Sierra Nevada its present great altitude. The summit peaks were raised to almost double their previous height, with Mount Lyell reaching more than 13,000 feet above sea level. At the same time, fracturing and faulting took place on an enormous scale. Owens Valley and other desert regions adjoining the range on the east and south subsided, or else suffered but slight uplifts as compared with the mountain block; and so the Sierra Nevada came to stand out in its present imposing form, with gentle westward slope, sharply defined crest, and abrupt eastward-facing escarpment. Strangely, the volcanic accompaniments of this great upheaval and inbreaking of the earth's crust were not extensive in the immediate vicinity. Though molten material forced its way up repeatedly through fractures in or near the zone of faulting, and also through cracks in the Sierra block, the resulting volcanic cones and lava flows were insignificant compared with those elsewhere in the Great Basin and northward in Oregon and Washington.

In the north-central Wasatch Mountains, the Wasatch fault broke and displaced an erosion surface of mid-Tertiary age. Most of the displacement was attained by the early Pleistocene (Eardley, 1944). See Fig. 31.7. Fresh scarps in the alluvium and across terminal moraines also attest post-Wisconsin movements.

Nolan believes that the best conclusion possible from present information is that block faulting probably began in places in early Oligocene time and has been more or less continuous ever since. Topographically expressed faults, however, probably date back only to late Pliocene or early Pleistocene, though earlier movements may have occurred along then.

Amargosa Chaos

An immensely disordered complex occurs in the Death Valley region which Noble (1941) has studied. See Fig. 31.8. In a centrally located district 10 miles square, called the Virgin Spring area, he finds the principal structure to be a flat thrust fault which originally followed approximately the contact of later Precambrian sediments and earlier Precambrian metamorphic rocks. On this thrust later Precambrian, Cambrian, and Tertiary rocks have moved relatively westward for an unknown distance. The rocks of the overthrust plate are broken into innumerable blocks and slices, which are thrust over one another to form an extremely complex mosaic. This assemblage of blocks is named the Amargosa chaos, and the flat fault upon which the chaos lies is named the Amargosa thrust. The chaos is divided into the Virgin Spring, Calico, and Jubilee facies. The Virgin Spring is characterized by blocks of late Precambrian and Cambrian dolomite, marble, sandstone, quartzite, shale, and slate. The Calico is made up almost wholly of Tertiary volcanic blocks, and the Jubilee contains a much larger proportion of poorly consolidated and broken-up material than the other two phases. The irregular blocks

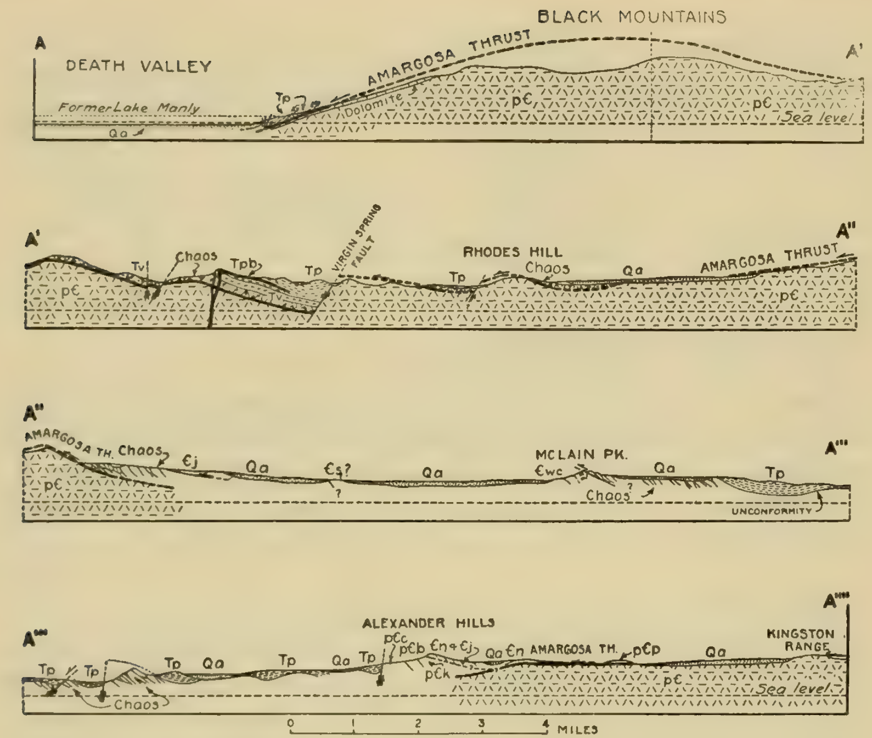


Fig. 31.8 Cross section of the southern Death Valley region. After Noble (Pl. 3, 1941). Tp, Pliocene (?) fanglomerate; Tv, undifferentiated volcanic rock; Ꞓwc, Wood Canyon formation (quartzite, shale, and fossiliferous limestone); Ꞓs, Sterling quartzite; Ꞓj, Johnnie formation (quartzite, shale, and dolomite); Ꞓn, Noonday dolomite; pꞒ, earlier Precambrian basement complex; pꞒk, Kingston Peak formation (conglomerate, quartzite, and shale); pꞒb, Beck Spring dolomite; pꞒc, Crystal Spring formation (quartzite, shale, and dolomite).

are granite, red Tertiary conglomerate, rhyolite, rhyolite tuff, porphyritic andesite, quartz latite porphyry, gypsiferous shale, fresh-water limestone and fanglomerate of Tertiary age, and various Precambrian and Cambrian rocks.

The Amargosa thrust and chaos are folded into several plunging anticlines of northwesterly trend, along whose crests the earlier Precambrian rocks below the thrusts are exposed. Lying unconformably upon the folded and eroded thrust sheets and chaos is the Funeral fanglomerate, probably of late Pliocene age, which consists of fanglomerates and basaltic lava flows. These rocks are deformed by folds and faults so recent that they are still reflected in the topography. The structure of Death Valley is thought to be a broad syncline modified by normal faulting. The Funeral fanglomerate is downfolded into this syncline and broken by step faults, downthrown toward the wide valley, along the east limb. These faults are, therefore, later than the Pliocene (?) Funeral fanglomerate. Very fresh scarps in Quaternary alluvium betray recent movement on them.

There is no evidence in this region of the Nevadan orogeny found to the west and north. There are, however, a number of large thrusts that bring older over younger Paleozoic rocks, which may represent the Laramide orogeny studied by Longwell (1928) and others farther east (Noble, 1941).

The Amargosa chaos terminates on the south against the east-west Garlock fault. Noble (1926) traced this fault eastward along the north side of the Avawatz Mountains, where it turns southward along their east side with reverse fault relations. Metamorphic rocks of probable Precambrian age are thrust against Tertiary beds (Nolan, 1943). A few miles farther east, Hewett (1928) has found remnants of a large horizontal thrust extending over an area of 30 square miles, along which Precambrian and lower Paleozoic rocks have overridden Miocene (?) sedimentary beds. The eastward movement of the thrust sheet is estimated to be at least 10 miles and may be as much as 20 or 25 miles.

The thrusting of late Tertiary age in southern California in the midst of the Basin and Range Province is most logically explained, it seems to the writer, as a gravity slide phenomenon incident to vertical uplift.

LATE CENOZOIC TRENCHES OF THE ROCKY MOUNTAINS

High Plateaus of Utah

Extending from the Coconino Plateau south of the Grand Canyon of the Colorado in Arizona northward to central Utah is a system of impressive fault scarps which bound a group of smaller plateaus and inter-

plateau valleys along the west edge of the Colorado Plateau. North of the state boundary, the assemblage is known as the High Plateaus of Utah, early described by Dutton (1880), and in northern Arizona the plateaus are known as the Kaibab, Kanab, and Shivwits.

In general, the faults and flexures block out ranges and intermontane valleys from the horizontal sediments of the Colorado Plateau, but toward the west the folded beds of the Laramide orogeny are involved. This is especially true in the Wasatch Plateau of central Utah and along the Hurricane fault of southern Utah and northern Arizona, previously described. The map of Fig. 31.9 shows the largest faults that have been attributed a post-Laramide age. Many small ones exist that are not shown, and even some major ones of which the age is uncertain or which have not been mapped as post-Laramide, may exist that are not shown. The Hurricane fault is illustrated in Figs. 20.21 and 20.22; the Sevier and Tushar faults, in Fig. 31.10.

A large volcanic field occurs in the central part of the High Plateaus and connects westward with other volcanic areas of the Great Basin. These are discussed in Chapter 36. They were mostly erupted immediately preceding the faulting.

An indication of the complexity of the volcanism, faulting, and erosion cycles of the region is revealed in Koons's (1945) work on the Hurricane and Toroweap faults just north of the Grand Canyon. The oldest eruptions of late Miocene or early Pliocene time preceded the earliest movements along the Hurricane fault and antedated the cutting of the Grand Canyon. They poured out on a large gently sloping pediment extending at least 16 miles north from the Colorado River. The main faulting then occurred, with displacements over 2000 feet at the Colorado River. The stream held its course, a new and lower pediment was eroded, and the region was brought approximately to its present configuration, with the Colorado River approximately as deep as now. The second eruption then occurred; they were local, and at the Toroweap fault filled the inner gorge to a height of 600 feet and perhaps 1200 feet. The lavas were entirely removed before later flows dammed the river again. These were subsequently also nearly all eroded away. Repeated movements along the Toroweap fault have occurred in late Pleistocene time, and in the very

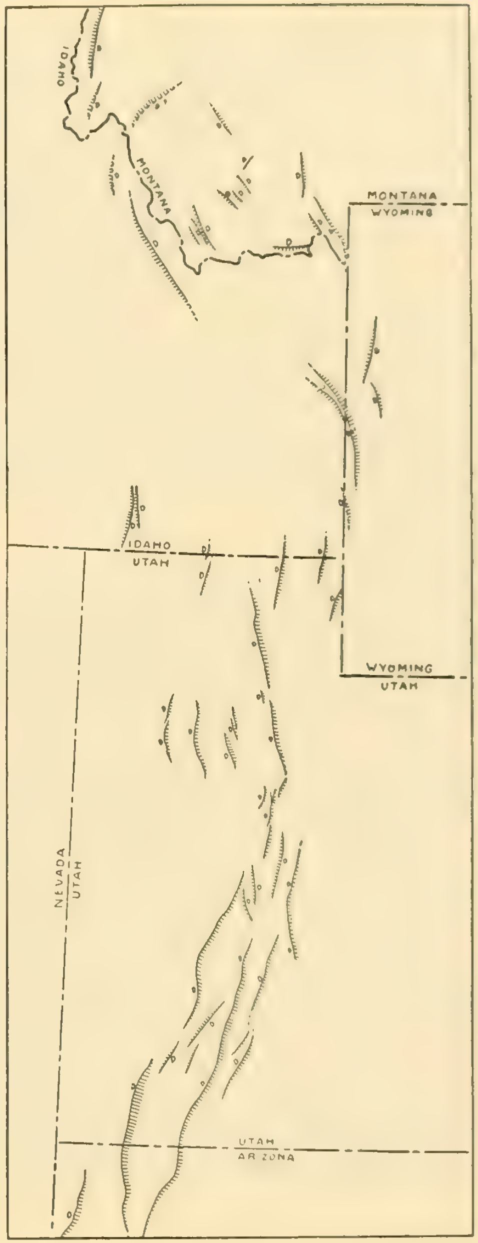


Fig. 31.9. Faults of the belt of great trenches in northern Arizona, Utah, Wyoming, Idaho, and southwestern Montana. Hachures are on the upthrown side. Only those faults are shown that have been fairly well demonstrated as late Cenozoic in age.

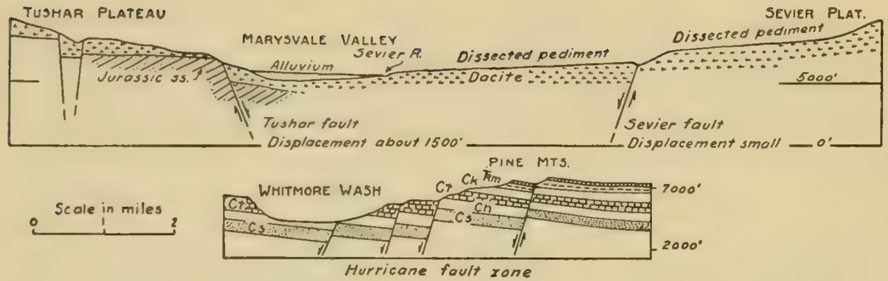


Fig. 31.10. Upper diagram: the Tushar and Sevier faults of the High Plateaus in Utah, after Eardley and Buetner, 1934.

Lower diagram: the Hurricane fault in Uinkaret plateau, northern Arizona, after Koons, 1945. ₸m, Moencopi fm; CK, Kaibab ls.; Ct, Toroweap fm.; Ch, Hermit sh.; Cc, Supai ss.

recent past renewed volcanic activity has formed a single, small cone and lava flow.

Wasatch Range

The late Cenozoic high-angle faulting along the west front of the Wasatch Range and the faults of the ranges immediately westward have already been described as part of the Basin and Range province. The belt of great trenches includes these faults.

Western Wyoming and Southeastern Idaho

Superposed on the Laramide structures of western Wyoming and southeastern Idaho are several northward-trending high-angle faults that have helped delineate and deepen the major intermontane valleys. Since the later structures parallel the earlier in northern Utah, southeastern Idaho, and southwestern Wyoming, the two have not been clearly distinguised; but toward the northern end of the belt in connection with the Snake River, Hoback, and Teton ranges, the Laramide structures veer northwestward, and the later high-angle faults cut across them at acute to right angles. A distinctive basin fill is also a result of the faulting, and helps distinguish the older from the younger.

A straight and youthful-appearing fault scarp occurs along the east side of Bear Lake in northern Utah and southeastern Idaho. It is responsible for the Bear Lake depression (Mansfield, 1927).

Star Valley in western Wyoming and its northward continuation in Grand Valley and Swan Valley between the Caribou and Snake River ranges is blocked out on one side and in places on both sides by faults of late Miocene and early Pliocene age. See cross section of Fig. 31.11.

An extensive graded surface had been eroded by middle Miocene time, and remnants of it still exist at elevations of 8500 to 9500 feet, especially in the Gros Ventre and Wind River ranges to the east. Blackwelder (1915) has called it the Union Pass surface. The main drainage lines of the present, except where affected by later faulting, had been established in and across the Laramide folds and thrust sheets by this time. Then the region was broadly uplifted, the streams rejuvenated, and the surface deeply dissected. The transverse and longitudinal canyons and valleys were eroded as deep as today and in the same position. These include the Snake River Canyon through the Snake River Range and the Hoback Canyon through the Hoback Range. Following the dissection of the Union Pass surface, normal faulting occurred as depicted in the series of diagrams of Fig. 31.12. In Grand Valley, west of the Snake River Range, the faulting and consequent deposition occurred in two episodes, and an unconformity was produced between two divisions of the valley fill. The sediments more than filled the graben and accumulated on the prefaulting surface to elevations above the fault scarp, and the canyons tributary to the graben that had previously been eroded in the Union Pass surface were flooded with debris. Toward the heads of these canyons, coarse material accumulated to elevations of 8500 feet. Volcanic activity accompanied the deposition of the valley fill, and much tuffaceous material was contributed to the deposits, and some thick sills split the basin beds. Then another cycle of erosion followed, and the Black Rock surface was cut at about 7500 feet. It was also a pediment that flanked the graben valley, and it beveled both the basin fill and the bedrock. The streams were again rejuvenated, perhaps several times, and the present valleys about 1000 below the Black Rock surface were eroded. The old faul

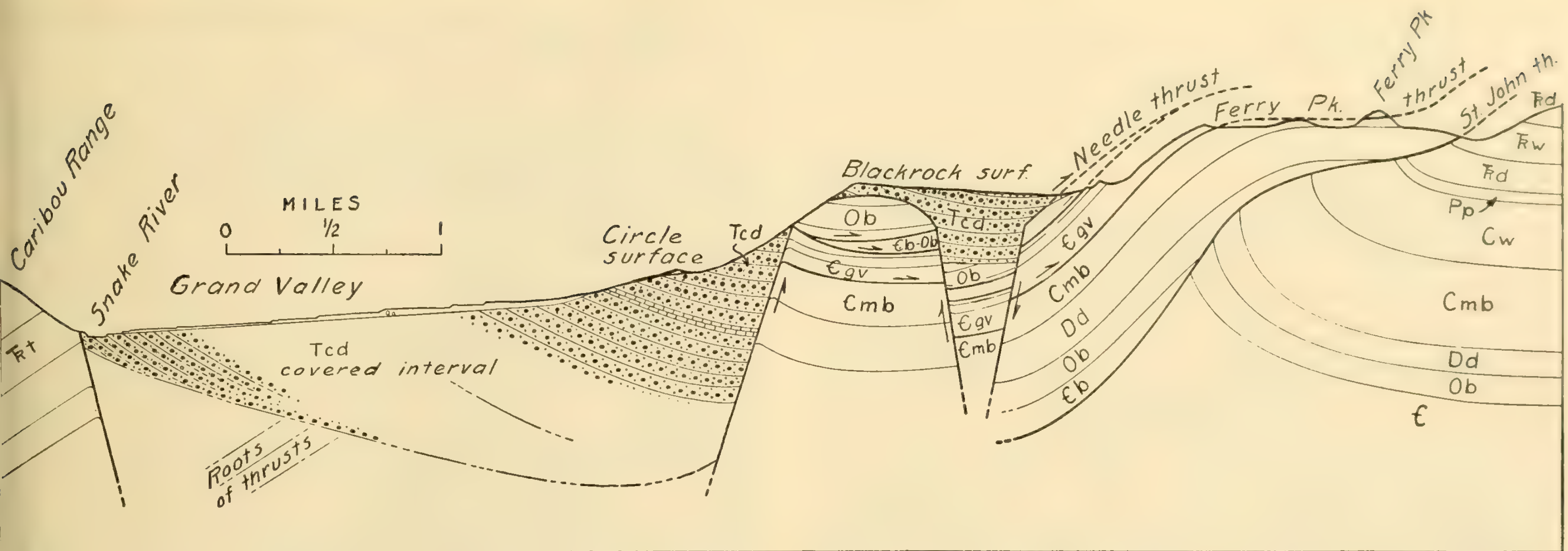


Fig. 31.11. Late Tertiary faulting near Alpine, Idaho, and Wyoming, and its relation to the Laramide structure. After Bayless, 1947. Ꞓgv, Gross Ventre formation; Ꞓb, Boysen limestone; Ob, Bighorn dolomite; Dd, Darby formation; Cmb, Madison and Brazer limestone; Cw, Wells formation; Pp, Phosphoria formation; Trd, Dinwoody formation; Trw, Woodside formation; Trt, Thaynes formation; Ted, Camp Davis conglomerate (upper Miocene or lower Pliocene).

carps that had been buried by the basin deposits were partly, but conpicuously, exhumed below the Black Rock surface. The one along the vest side of Grand Valley has all the physiographic features of a youthful ault scarp, yet is a fault-line scarp.

Jackson Hole, between the Teton and Gros Ventre ranges, is the result f downdropping along the Teton fault (Horberg, 1938; Love, 1956a). he Union Pass surface is believed to have been broken and rotated so hat it passes below the valley fill on the Gros Ventre side, and has been levated and tilted westward on the Teton side. The basin deposits, largely onglomerates, tuffs, and lavas, may have been folded somewhat after leposition, but this aspect of the history is not clear.

The discordant relations of the Grand Valley, Hoback, and Teton faults to the Laramide structures in map view are shown in Fig. 31.13.

Southwestern Montana and Central Idaho

Fresh fault scarplets occur along the west base of the Madison and the Tendoy ranges of southwestern Montana, and major fault scarps occur along the east faces of the Blacktail and Ruby ranges, and the northwest face of the Bitterroot Range.

The northeast face of the Lemhi Range in Idaho is thought to be, in part at least, a fault scarp. There may be others, but these are the only ones that the writer has seen. Although not yet studied in detail, these

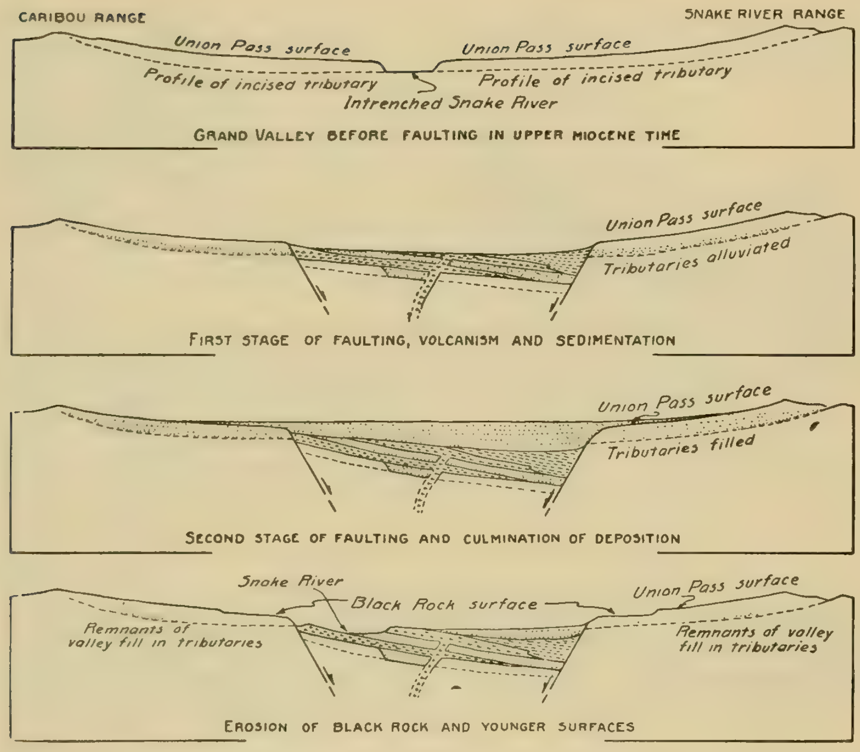


Fig. 31.12. Idealized diagrams showing the late Cenozoic evolution of the Grand Valley trench in Wyoming and Idaho.

mountain fronts appear surprisingly like the classical Wasatch scarp in Utah. Basin beds are widespread in the large intermontane valleys, and in part were deposited before the faulting and have been displaced by it, but in part are a direct consequence of it. In the erosion that followed the faulting, the basin beds have been stripped away in places from bedrock against which they had been faulted or in other places deposited, and fault-line scarps have formed, as in Fig. 31.12. The basin beds in which fossils have been found are upper Eocene, middle Oligocene, lower Miocene, and uppermost Miocene or lower Pliocene, and have a large tuffaceous and volcanic ash content, and even sills or lava flows in places. The Tertiary history is reviewed under the heading "Southwestern Montana," in Chapter 22. See also Fig. 31.14.

Northwestern Montana, British Columbia, and the Yukon

The Rocky Mountain Trench of British Columbia is described in Chapters 21 and 33. It continues the zones of great trenches to the Yukon and probably to Alaska.

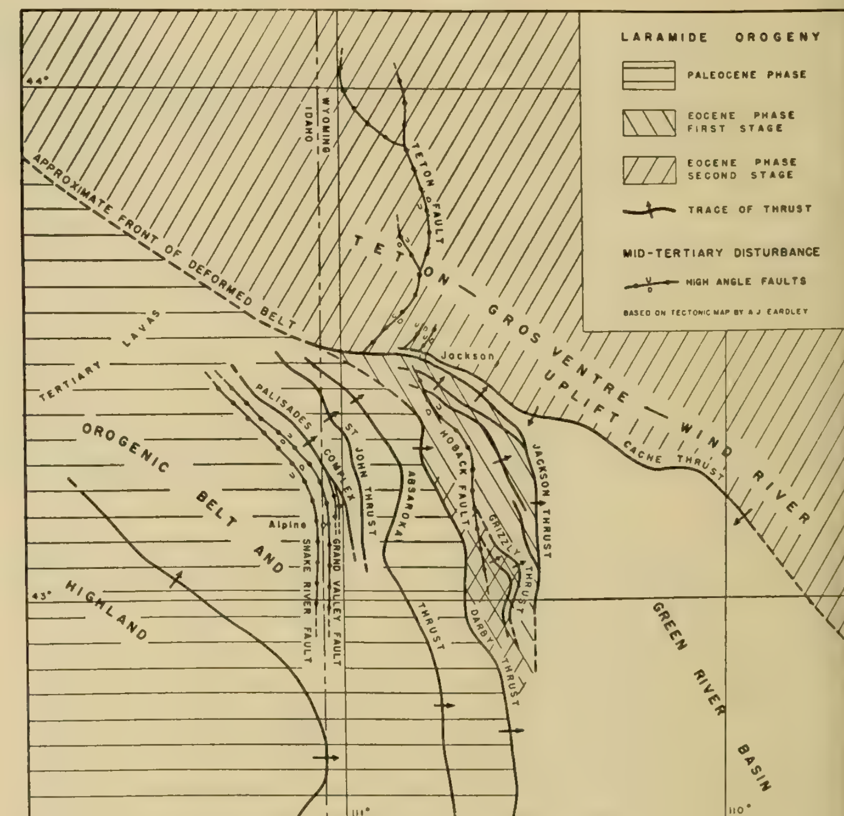


Fig. 31.13. Relation of late Tertiary faulting to the Laramide elements in northwestern Wyoming and eastern Idaho. After Bayless, 1947.

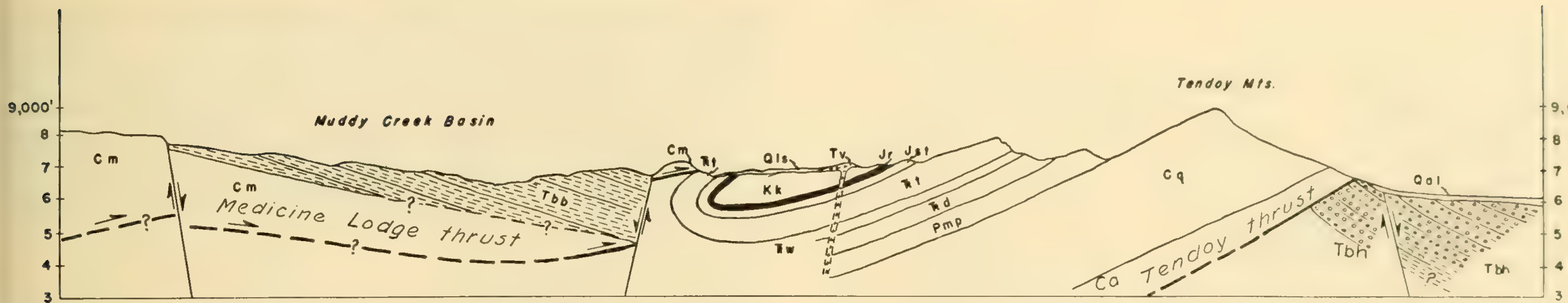


Fig. 31.14. Cross section of the Tendoy Mountains; Cm, Madison ls.; Ca, Amsden fm.; Cq, Quadrant quartzite; Pmp, Phosphoria fm.; TRd, Dinwoody fm.; TRw, Woodside fm.; TRt, Thaynes fm.; Jst, Sawtooth fm.; Jr, Rierdon fm.; Kk, Kootenay fm.; Trr, Red Rock conglomerate (Paleocene ?); Tbb, Muddy Creek basin beds.

Seismicity in the Trench Zone

After the past pages on the zone of great trenches that extends from Arizona to the Yukon had been written, attention was called to the earthquake maps of Woolard (see Fig. 31.15). The concentration of major shocks in the zone of trenches is striking. The coincidence not only supports the existence of the zone of faults but also indicates that a number of them are still active.

GEOPHYSICAL EVIDENCE

Gravity and Seismic Surveys

The fill of the down-faulted basins in the Basin and Range provinces lends itself to analysis particularly by gravity surveys. Since the alluvium has lighter density than the lithified bedrock, the magnitude of the gravity anomaly can be related to the depth of fill, and this becomes a measure of the magnitude of faulting. Also, faults concealed beneath the alluvium may be detected, and new light is shed on the fault pattern. The computed cross sections on the basis of gravity surveys have been checked by seismic surveys across the valleys.

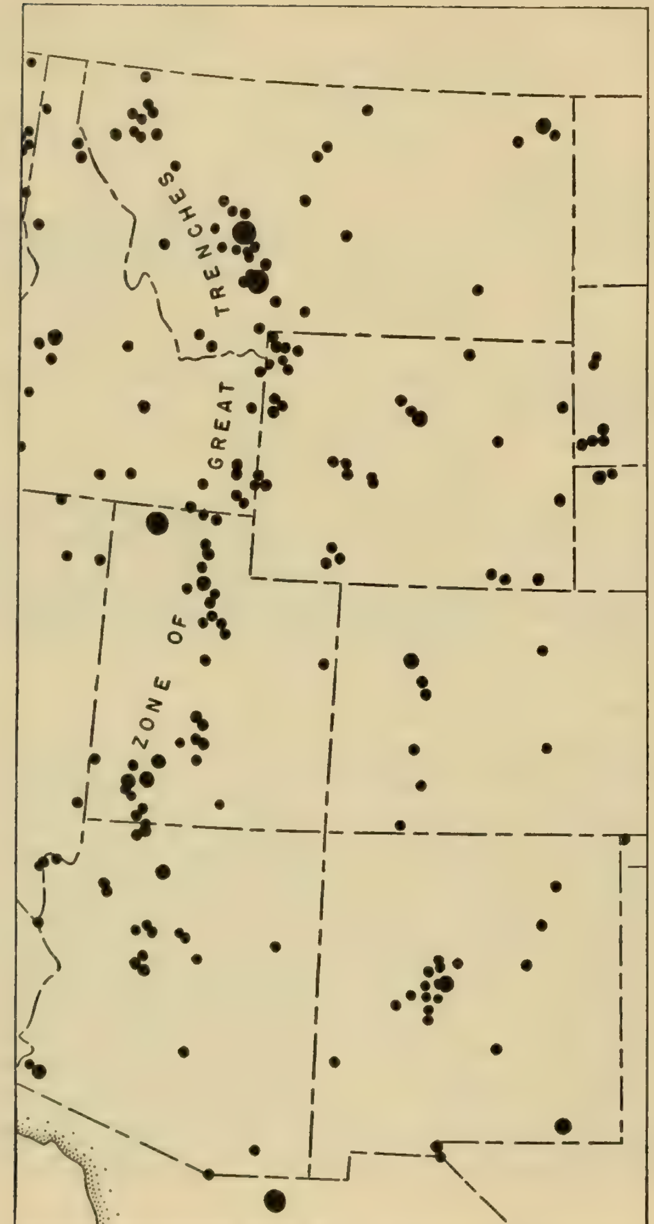
Recent earthquakes in the Great Basin have been studied seismically and the results add to our concepts of Basin and Range structure.

Fault Patterns

Two kinds of patterns appear at present to exist. The one consists of subparallel faults which define graben, horsts, and tilted blocks, and the other of faults in semicircular or polygonal form which bound completely or nearly completely downfaulted blocks. The two are illustrated in Fig. 31.16 of the Owens Lake–Mono Lake region of California.

Mono Lake Basin

Mono Lake is in a somewhat triangular-shaped basin about 15 miles in length at the eastern foot of the Sierra Nevada. As a result of gravity and seismic studies Pakiser *et al.* (1960) conclude that nearly vertical faults bound the triangular-shaped block, and that it has subsided 18,000 ± 5000 feet and has received about 300 ± 100 cubic miles of light clastic sediments and volcanic material of Cenozoic age. The nature of the gravity profile and the interpreted geologic section on the northwest side are shown in Fig. 31.17. A section across the entire basin is given in Fig. 31.18. It will be seen that the basin fill is divided into layered deposits, a lower thick one of relatively high velocity (7800–10,800 feet per second) and an upper thin one (2000 feet) of low velocity (5500–6200 feet per second). The recent deposits have not been displaced by faulting and conceal the buried faults. The lower deposit is believed to be mostly Tertiary volcanic



material, and the cause of subsidence of the pluglike block to be due to the relief of pressure from below by the movement away of magma in a supporting chamber. The magma is presumed to have found escape at the surface, but only part of the extrusives accumulated in the subsiding basin. The magmatism and pluglike faulting are believed to be related to the general tectonic framework of deforming forces of the Basin and Range provinces. The nature of the relation will be considered in following paragraphs.

Region West of Wasatch Range

Cook and Berg (1957 and 1961) report on an extensive gravity survey in Salt Lake and Utah counties where they made 1100 observations over an area of 5000 square miles. Steep gravity gradients reveal buried faults unrecognized by surface geologic surfaces, and although the downfaulted valley block between the Wasatch and the ranges on the west was known to contain over 2000 feet of unconsolidated or semiconsolidated sediments a deep inner trough was discerned which with a number of irregularities extends north-south for over 100 miles. "Several large fragments . . . have apparently dropped deeper than the other fragments, as if slipping into a great crevasse."

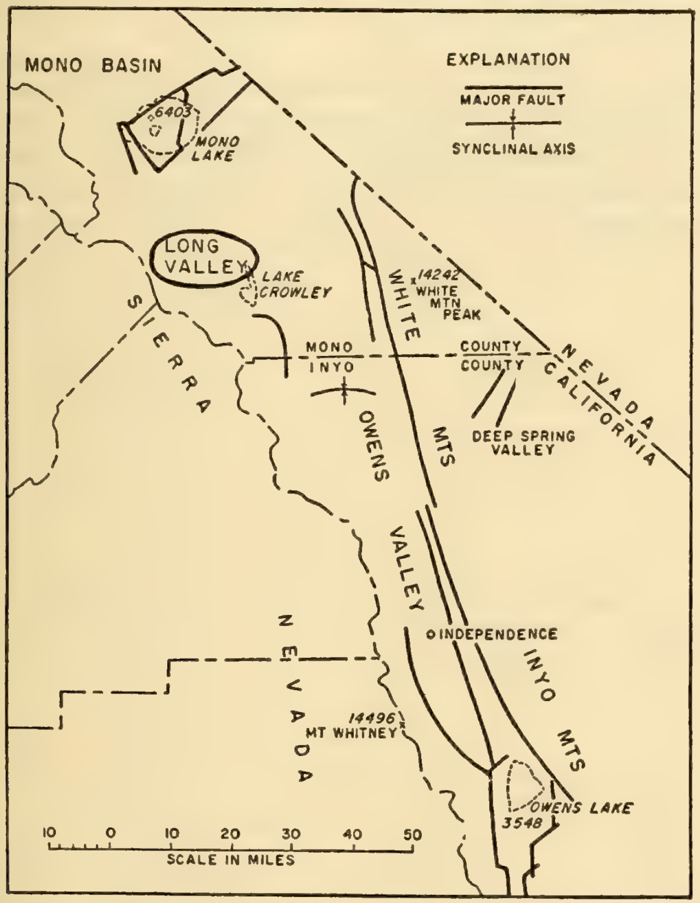
Fallon–Austin Earthquake Area

A major earthquake occurred in the Dixie Valley–Fairview Valley area of west-central Nevada in 1954, and fresh scarps were formed. Their pattern is shown in Fig. 31.19. The faulting is most advantageous to study because a first order triangulation net and a first order line of levels had been established across the area before the movements. The stations were reoccupied and the amount of vertical and horizontal movement accurately determined. A vertical displacement of 7 feet occurred in Dixie Valley and also 7 feet where the fault is in bedrock east of Fairview Peak. The arrows of Fig. 31.19 indicate the horizontal extension that occurred and which averages about 5 feet in magnitude in a northwesterly direction.

Fig. 31.15. Earthquake epicenters of the Rocky Mountain region showing coincidence of zone of concentrated seismic activity and the belt of trenches. Taken from map compiled by G. P. Woolard from U.S.C. & G.S. reports.

There was no displacement of points 40 miles west and east of the fault zone.

A gravity profile across Fairview Valley and the interpreted geology are shown in Fig. 31.20 (Thompson, 1959). The valley fill is about 1 mile thick, and the topographic relief of the adjacent range is about 1 mile, so that Thompson concludes a total cumulative vertical displacement of



ig. 31.16. Index map of Basin and Range faults in the Mono Lake—Owens Lake area immediately east of the Sierra Nevada. Reproduced from Pakiser et al., 1960.

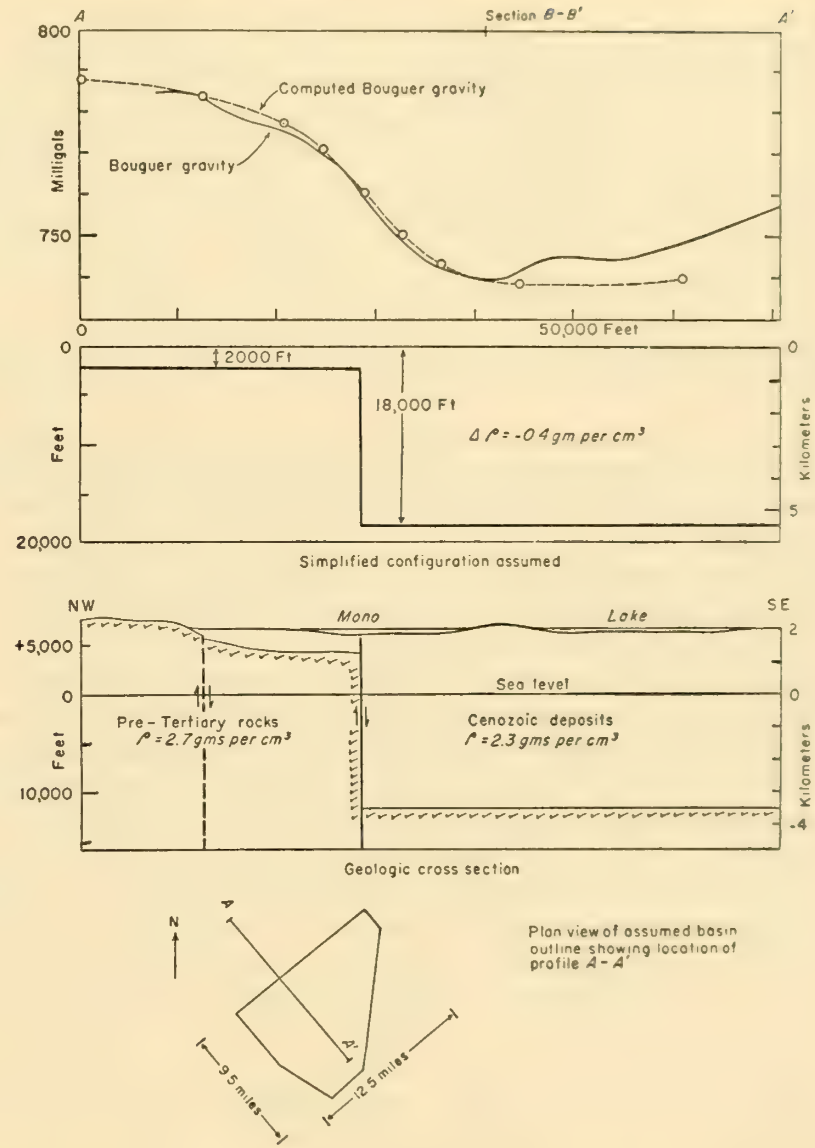


Fig. 31.17. Mono Lake basin interpreted from gravity profile. Reproduced from Pakiser et al., 1960.

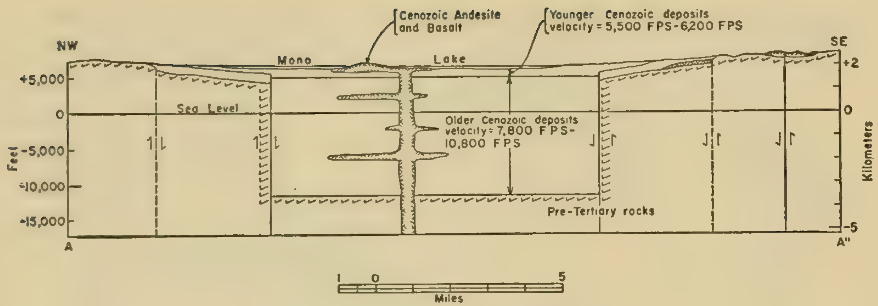


Fig. 31.18. Generalized geologic section across Mono basin. Reproduced from Pakiser *et al.*, 1960.

about 2 miles has occurred since the inception of faulting, which he assumes here was in the Miocene. If the basin is bounded by normal faults considerable distention of the crust must have occurred over the course of movement. If the basin is bounded by faults dipping 60 degrees (lower diagram, Fig. 31.20), the extension normal to the strike amounts to about a mile on each side of the basin or a total extension of 2 miles. If the faults dip 70 degrees, the extension amounts to about 1½ miles.

The location of the focal depths probably reveals the depth to which faulting extended. Two earthquakes occurred 4 minutes apart in time and 35 miles apart in distance. The southern Fairview Peak focal depth was determined by Romney (1957) to be 15 kilometers below the surface, and the northern Dixie Valley one to be 40 kilometers. Also a close correspondence of dip and direction of motion at the surface was found to obtain at the 15-kilometer focus. These points lead Romney to believe that the fault fracture extended to a depth greater than 15 kilometers. The even greater depth of the northern focus supports the conclusion that the entire crust to the Moho discontinuity is possibly affected. Two possible fault structures are shown in Fig. 31.20, with the one on the right coming closest to fitting the facts (Thompson, 1959).

The amount and rate of distention of the entire Basin and Range province are estimated by Thompson as follows:

> The data indicate that the region of Dixie and Fairview Valleys has been distended in a nearly east-west direction about a mile and a half. If we assume that each of the principal basins between the Sierra Nevada and the Wasatch Mountains has been deformed this much on the average, the total distention amounts to 30 miles or 5 pct. And if the deformation took place in the last 15 million years, as suggested by the geologic history (deformation of Miocene-Pliocene and younger rocks), the rate is 2 mi/million years or only 1 ft/century. The rate of extension indicated by several fault movements within historic times appears to be at least 1 ft/century. The faults lie in a north-south belt about 250 mi long. For at least this distance the data are consistent with an extension of 1 ft or more in the last hundred years. Prehistoric Quaternary faults are also numerous; they strongly suggest that the historic rate of deformation is not abnormally high.

Tilted blocks, which are characteristic of large parts of the Great Basin, may or may not be the result of extension of the crust. If they are an expression of tension then the general level of the surface is depressed and the crust thinned. Since the Great Basin appears from other geological considerations to be a depressed region, the tilted blocks

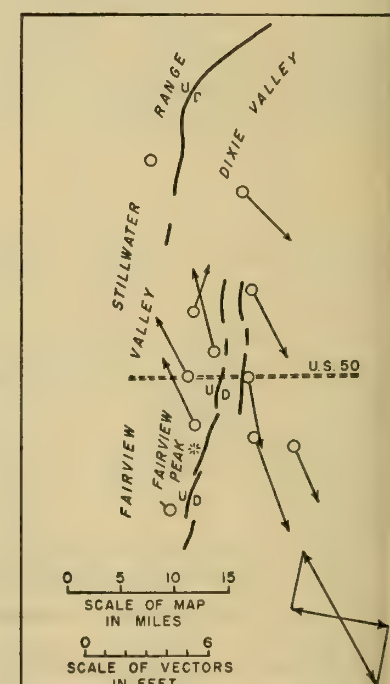
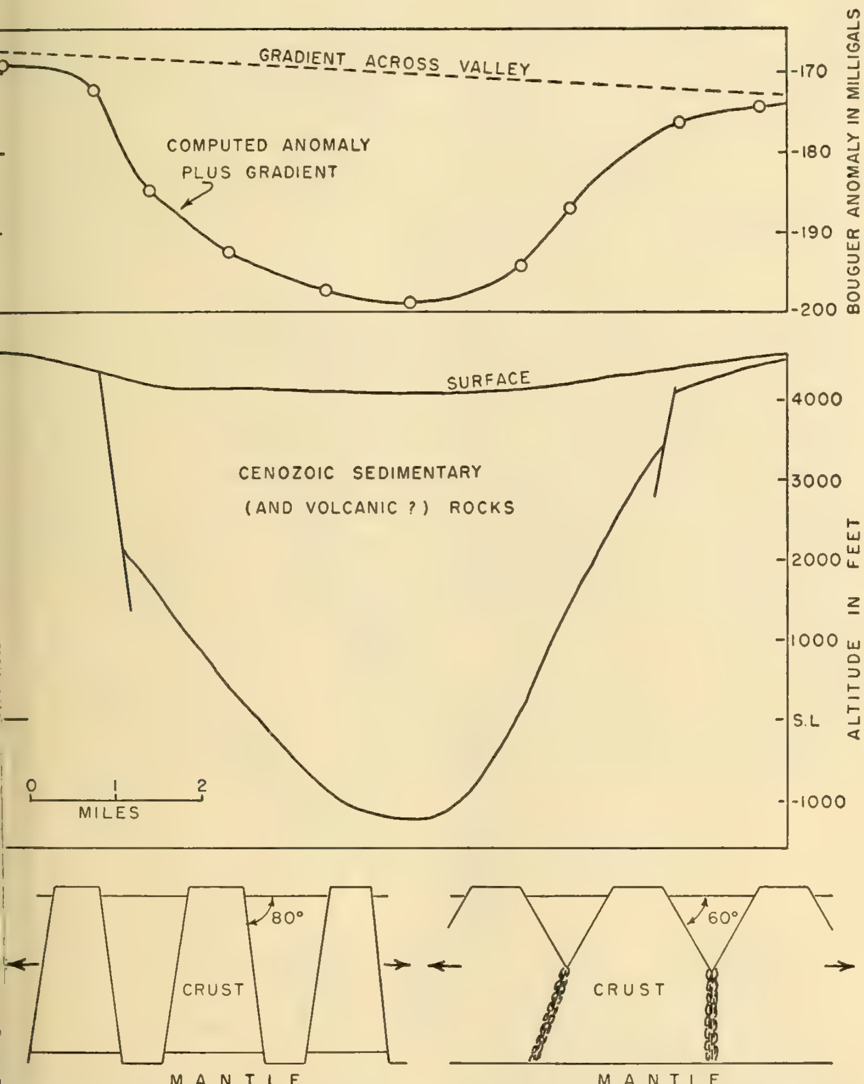


Fig. 31.19. Horizontal movements in the Fairview and Dixie valleys earthquake. After Thompson, 1959.



ɟ. 31.20. Gravity profile and section across Fairview Valley. Also alternate interpretations of ulting of crust under extending forces. From Thompson, 1959.

will be considered tensional features as well as the graben blocks.

If the crust has been extended some 30 miles between the Sierra Nevada and Wasatch Mountains, then our understanding of the penetration of magma into and through it comes into better focus. In Chapter 33 it is suggested that the large volumes of quartz monzonite magma originated in the base of the silicic (granitic) layer of the crust at depths of 10 to 20 kilometers, and we can see that the tensional fractures illustrated by Thompson in Fig. 31.20 would penetrate such magma chambers and conduct the magma upward. From this point of view both the block faulting and magmatism are the result of the tensional tectonism, and only in the local examples of pluglike basin subsidence should we conclude that the evacuation of a magma chamber is the direct cause of the faulting.

We are led to speculate that fractures have penetrated to the basaltic subcrust in Oregon and Washington to conduct the olivine and tholeiitic magmas to the surface.

EXPLORING TENSIONAL TECTONISM IN WESTERN NORTH AMERICA

The theory of expansion of the Basin and Range province in late Cenozoic time in the magnitude of 30 miles piques one's curiosity to consider the entire framework of movements in western North America. The strike-slip movement along the San Andreas system and the postulated extension of the Basin and Range province with its components of horizontal movement should be related. Figure 31.21 has been prepared to show the directions of fault traces and the horizontal movement on the San Andreas. Only a few of the faults of the Great Basin are shown such as to indicate the direction of tensional forces that must be entertained.

Figure 31.22 is a diagrammatic map which resolves in bold strokes the distention cracks and horizontal movements of the crust previously postulated. The expansion fractures of the Basin and Range province are distributed across the entire basin, but for purposes of illustration are concentrated along the eastern and western margins. The width of the lines represents the approximate amount of postulated expansion. The main pulling away appears to have been in a west-northwesterly direction

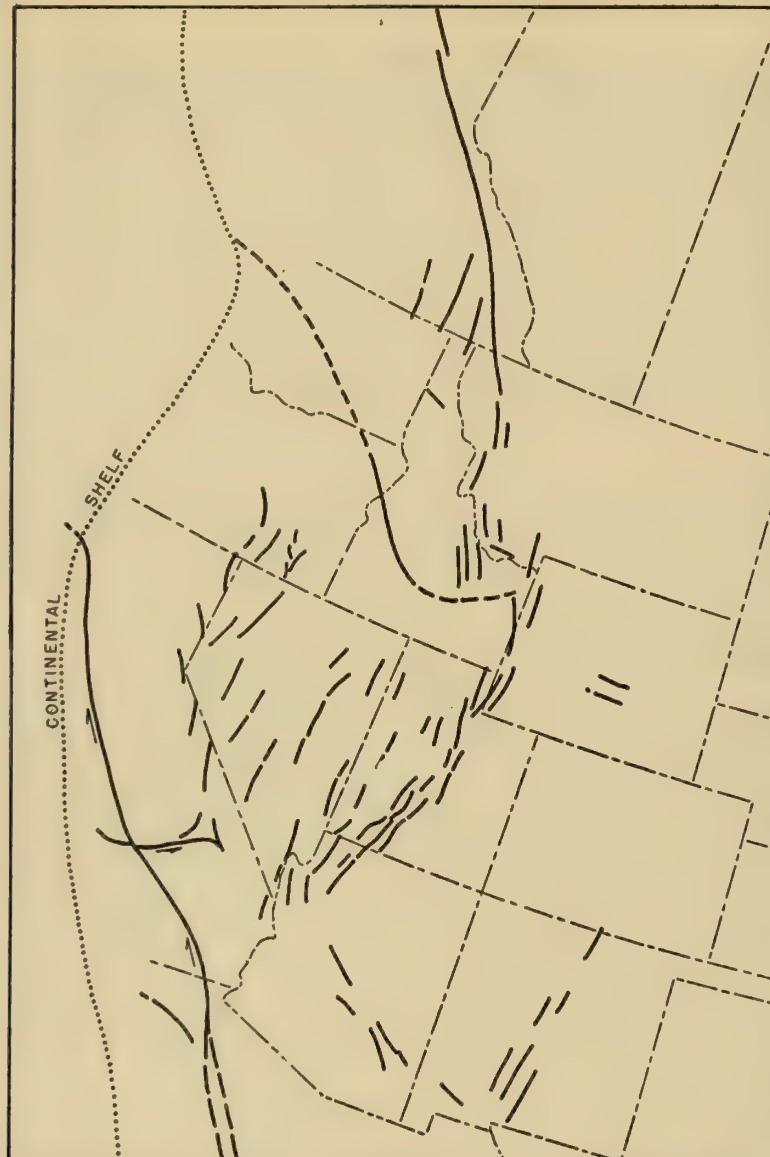


Fig. 31.21. Framework of late Cenozoic fault systems of western United States.

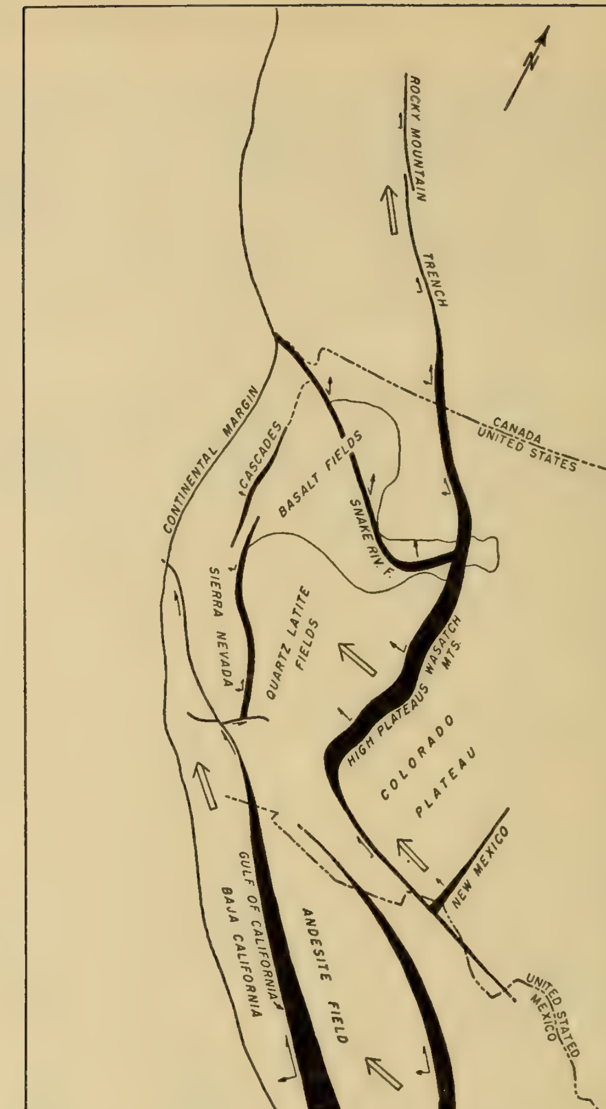


Fig. 31.22. Exploring the concept of extension and drift affecting western North America. Black lanes represent amount of expansion as if localized along a few separations. Except for the Gulf of California the extension is distributed in a number of separations across the entire Basin and Range province. Small arrows represent apparent vectors of movement. Large arrows the apparent resultant direction of movement.

with a strong northwesterly component in central California keeping the Coast Ranges block snug against the adjacent continental mass. Perhaps the same is true of the Rocky Mountain Trench. Some separation and also horizontal displacement have been postulated for the Rocky Mountain Trench. This movement is possible when the Snake River fault is considered to be one of considerable separation (Chapter 36).

The drifting away from the continent of Baja California as well as a northwesterly gliding movement seems substantially demonstrated. See Chapter 29.

A major strike-slip fault is postulated across south-central Arizona at the south margin of the Colorado Plateau. Southern Arizona remained 5000–8000 feet below the Colorado Plateau after vertical adjustments occurred in late Cenozoic time, and is generally considered to be a block-faulted region, although not so clearly as the Great Basin in western Utah and Nevada. A few alluvial-filled valleys parallel the grand escarpment and support the concept of down-dropping along major faults. However, a master horizontal couple as indicated on the map of Fig. 31.22 has not been recognized or postulated, as far as the writer knows. This then, is a very speculative element of the framework of movements illustrated on the map.

The rifting of central New Mexico finds a compatible place in the framework. The Sonoran-Chihuahua basin and range region is poorly understood, and the illustration of considerable distention there is hardly more than a guess.

SEISMIC VELOCITY LAYERS IN THE EASTERN GREAT BASIN

Seismic Layers

The recognition of a crustal layer with a velocity of 7.5 ± kilometers per second in several areas of the western United States and Canada comes as a very significant find and perhaps a key to tectonism there. The work of Berg *et al.* (1960) in the eastern Great Basin, Press (1960) in the California–Nevada region, and the summary article by Diment (1961) should be referred to. The seismic velocity layers recognized to date are portrayed in Fig. 31.23.

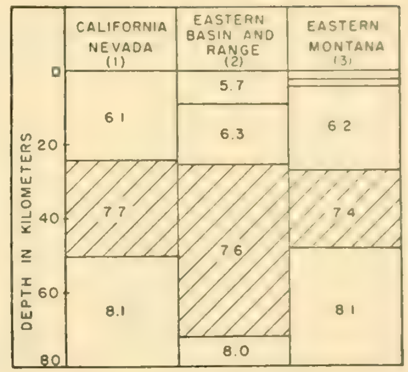


Fig. 31.23. Seimic velocity layers in western United States Velocities in kilometers per second (1) Press, 1960; (2) Berg *et al.*, 1960; (3) Meyer *et al.*, 1960. Refer to Fig. 38.1.

Geologic Requirements

In attempting to interpret the constitution of the seismic layers the following geologic requirements should be kept in mind.

1. The Great Basin has been distended about 30 miles (50 kilometers) in the last 15 m.y. A strong horizontal coupling along the Pacific margin is evident, with the Pacific facing blocks moving to the northwest.
2. The Great Basin has been elevated during the same time 1–1½ kilometers.
3. The High Plateaus of Utah and the Sierra Nevada have been elevated 2–3 kilometers during the same time.
4. The Colorado Plateau has been elevated 2–2½ kilometers during the same time.
5. Silicic lavas have been poured out over most of the Great Basin in amounts equal to a layer 1–2 kilometers thick since early Oligocene time. This material must have come from the melting of a portion of the silicic crystalline mantle. See Chapter 36.
6. Equal amounts of basalt (viz., the Columbia basalt field) have flowed to the surface from a source probably immediately below the crystalline basement, and in the Great Basin the basalt reservoir has been tapped from time to time during the general acidic lava eruptive cycle.

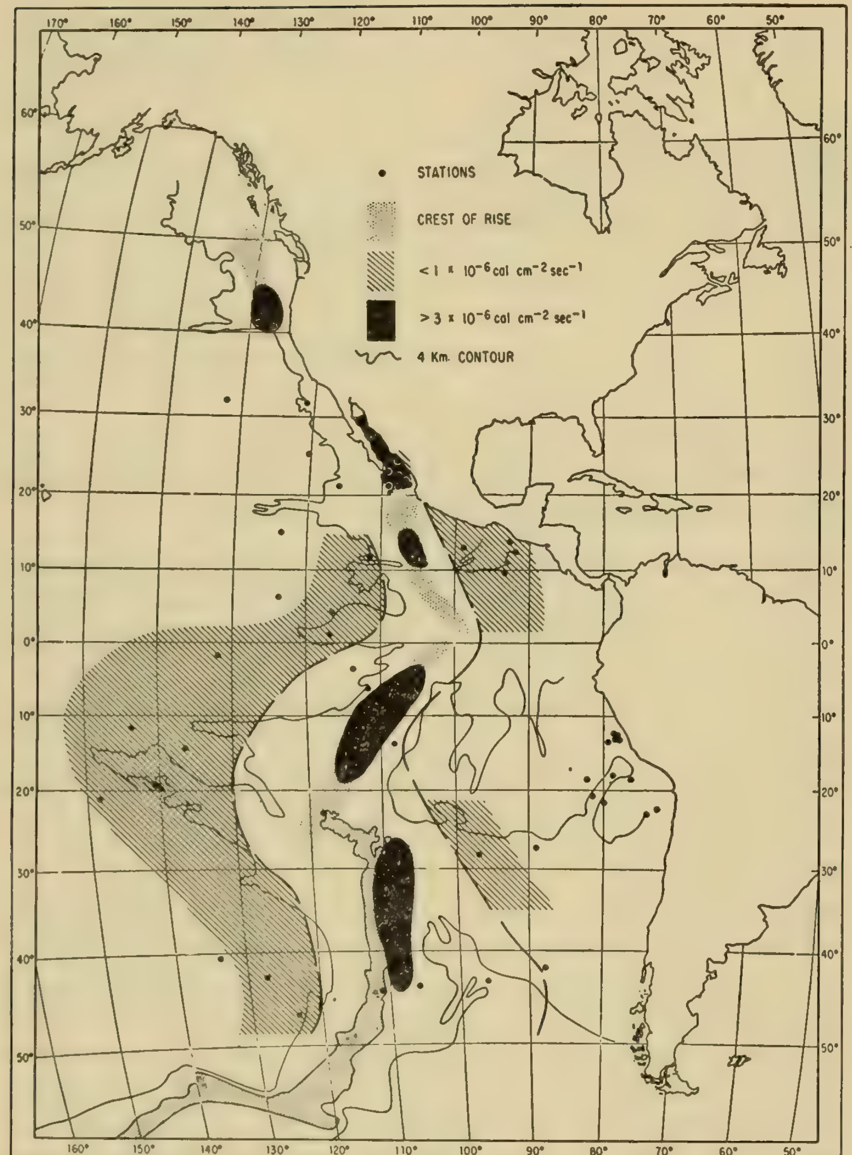


Fig. 31.24. East Pacific Rise and pattern of heat flow. Reproduced from Menard, 1960.

Interpretations

The East Pacific Rise of the ocean floor has been considered by Menard (1960) to extend to the Gulf of California and hence under the western part of the continent of North America appearing in the Pacific again off Oregon, Washington, and British Columbia. See Fig. 31.24.

> The puzzling slope between California and Hawaii is the west flank of the rise. . . . Where the crest and east flank of the rise intersect Mexico are found the plateau of Mexico, the Colorado Plateau, and the Basin Ranges comprising a topographic bulge of the continent comparable in scale to the bulge of the sea floor.

Cook (1961) follows Menard in projecting the East Pacific Rise under the continent, and assigns the broad uplift to the development of the 7.4–7.7-kilometer-per-second velocity layer under it. In fact, he believes from still incomplete data that the oceanic rises of the Pacific, Atlantic, and Indian oceans with their accompanying rift systems and volcanism are due to the uplift of the crust as the 7.5 layer develops. He calls it the mantle-crust mix layer, and regards it as a change from eclogite to basalt with attendant expansion.

The views of Menard and Cook related to the western United States lead to many thoughts which will only be summarized here. First, the Late Cenozoic uplift should be considered. Approximate uplift contours are shown in Fig. 31.25. They are admittedly approximate, and in the Great Basin represent an average of the uplift of the Tertiary deposits in the valley blocks and the uplift of the mountain blocks. From the picture presented the Snake River downwarp and associated Columbia basalt region may represent a transverse break in the continuity of the 7.5-km/sec layer from south to north. The Colorado Plateau has been uplifted more than the Great Basin, and it has generally been considered that the Great Basin is one of collapse or subsidence in relation to the Plateau, although in relation to sea level, both have been uplifted. It will be very interesting to see what the relative heat-flow measurements will indicate as to the central part of the rise over the 7.5 layer. None has been made yet. Cook seems to infer that the zone of Great Trenches and accompanying seismicity is the central rift zone of the rise. Fig. 31.15.

In Chapter 36, the igneous rocks of the western United States are re-

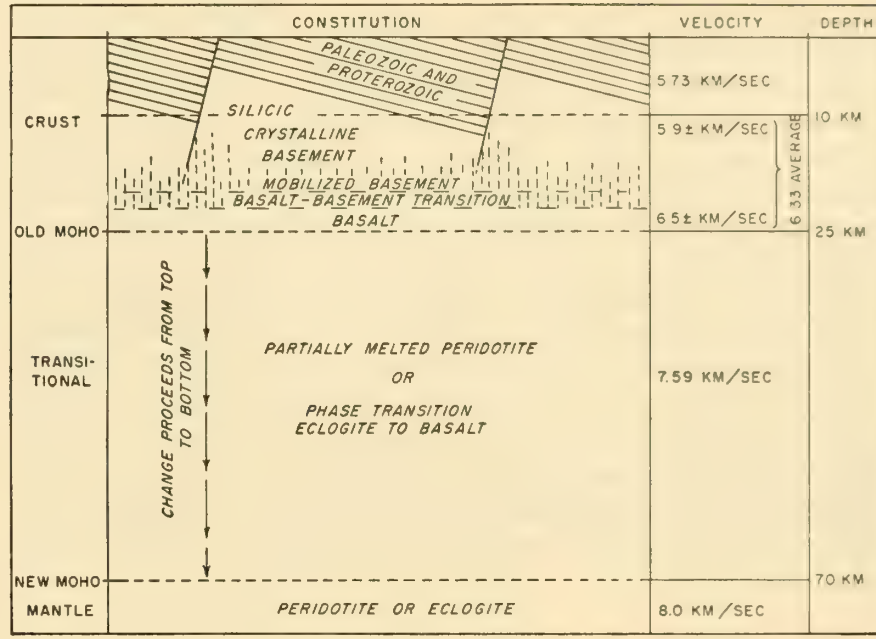


Fig. 31.26. Postulated constitution of velocity layers under eastern part of Great Basin.

viewed. These must certainly be considered in visualizing the constitution of the crustal layers and the role of the 7.5 layer in tectonism. The writer's ideas of the arrangement in the eastern part of the Great Basin are shown in Fig. 31.26, and are discussed as follows.

The mantle is regarded as either peridotite or eclogite. If the 7.5 layer is a transition layer, as seems necessary from its seismic velocities, then if the mantle is peridotite, the transition layer would be one of peridotite and its early melt product, basalt. If eclogite, then basalt or gabbro would result as a phase transition. In either case, Cook's name, mantle-crust mix, would be suitable. The writer favors the peridotite-basalt mix, because he sees in it a means of bringing molten basalt in large quantities upward

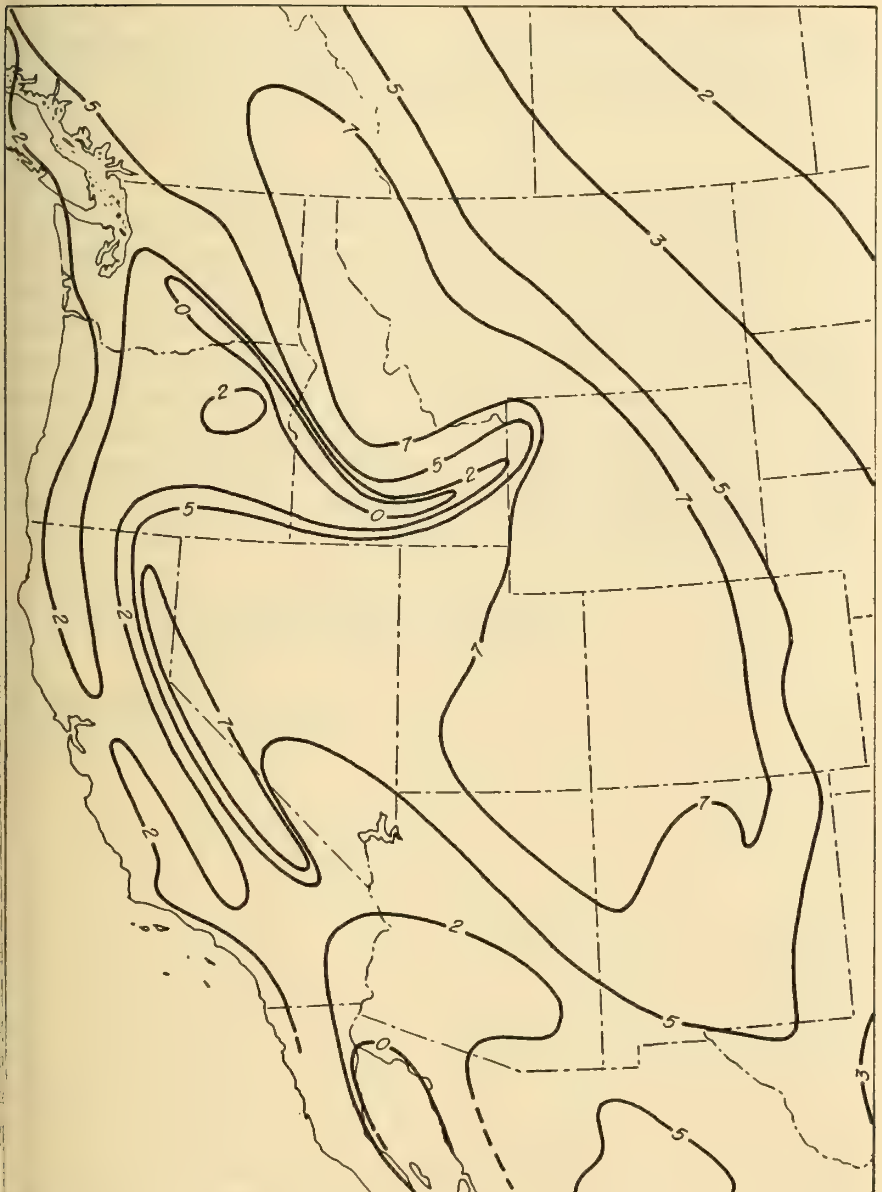


Fig. 31.25. Late Cenozoic uplift in western United States. An attempt is made to portray the broad vertical movements of the silicic crystalline basement layer. Contours in thousands of feet.

to the base of the silicic crystalline basement layer. This is necessary to feed basalt to the surface in ways listed in the geologic requirements previously mentioned. The basalt layer is visualized as growing in thickness as the molten basalt from below rises and is added to it. In case of tensional fractures in the crust which reach downward through the basement, the basalt reservoir is tapped, and fissure flows result. When eclogite changes to solid basalt through polymorphic phase transitions, much heat is consumed in the process, and unless considerably more is generated in the mantle or the basalt so formed, none converts to liquid basalt.

The heat of the liquid basalt which has risen to the base of the silicic crustal layer mobilizes, if not melts, a considerable amount of it; and it is this silicic magma which is postulated to have erupted at the surface to form the voluminous silicic flows of the Great Basin and the alkalic igneous rocks of the shelf province (Chapter 33).

Menard postulates a convection current rising under the East Pacific Rise and flowing westward under the crust. The drag of this current creates tensional block fault features in the central zone of uplift, it translates the adjacent crust westward, and in the region of downward plunge of the current, compressional structures are formed. He has difficulty, however, fitting the San Andreas fault into the convection current hypothesis.

Reference to Figs. 31.21, 31.22, 31.25, and 32.15 should convince one that the cause of late Cenozoic tectonism must be complex, and more is involved than westward movement of the convection cell. In addition to the San Andreas fault with large strike-slip movement to the northwest, there is the Snake River fault which appears to separate the western Cordillera into two distant segments. A drift of the crust to the northwest with extension to the west-northwest is fairly clearly indicated. Besides variations of convection circulation and expansion of the mantle to accomplish these movements of the crust, there is need to consider the (as yet intangible) forces presumed to cause polar migration, drift, and rotation of the continents. The pattern suggests such forces to the writer.

32.

PACIFIC SUBMARINE PROVINCES

DISCOVERY OF STRONG SUBMARINE RELIEF

It was current opinion until 1925 that the ocean floors were montonous plains. The continental shelves above the floor and the great deeps below the floor were known, but not their details. The technique of echo sounding was successfully introduced in 1925 by the U.S. Coast and Geodetic Survey, and since then remarkable progress in mapping the floor of both the Pacific and Atlantic oceans has been made. Many thousands of miles of traverses have been run, and with progressively more accurate means of location available the contouring has become more accurate and the topography better known. The Gulf of Alaska was explored before 1940, and instead of a featureless floor a number of bold seamounts were discovered. The most detailed early survey was off the coast of southern California, where basins, banks, ridges, and escarpments of comparable size to those on the adjacent land were indicated.

In addition to many seamounts in the northeastern Pacific, various ridges, depressions, and trenches were discovered, and by 1955, the length of sounding lines to show the extent and some of the details of these features had reached about 80,000 miles (Menard, 1956). This work was done chiefly aboard ships of the Navy Electronics Laboratory and the Scripps Institution of Oceanography. Several expeditions each year since 1955 continue to add to an ever amazing picture of the Pacific ocean floor.

Study of the submarine topography is pertinent to an understanding of the deformation of the oceanic crust, and most interpretations to date have been made from the relief features. Valuable supplementary information has come from seismic and gravitational surveys, and most recently from extensive magnetic intensity surveys.

SUBMARINE PROVINCES

Basins, Banks, and Ridges off California

The submarine topography for 150 miles off the southern California shore is one of basins, banks, and ridges comparable with that of the adjacent land. Shepard (Shepard and Emery, 1941) calls it the continental borderland. See Figs. 32.1 and 32.2.

In this borderland are eleven basins which would contain large lakes if the land became emergent. Some of them would cover 1000 square miles and would range up to 2880 feet deep. The basins are roughly oval and elongated northwesterly. Their walls are generally steep, long, and straight, but are gashed by a few valleys. However, abrupt changes in direction exist. The basin floors are very flat, and do not possess the piedmont slopes of their land counterparts in southern California and Nevada. The general elevation of the basins and their overflow sills becomes greater to the southeast (Shepard and Emery, 1941).

The elevations on the continental borderland are numerous and diverse.

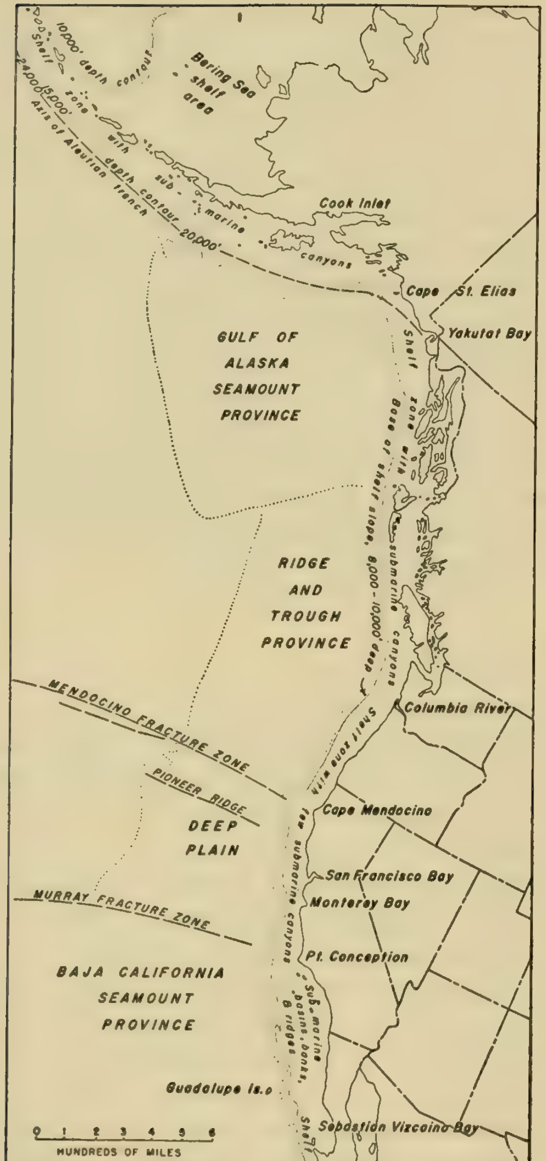


Fig. 32.1. Pacific submarine relief provinces off North America. After Menard, 1955.

The higher elements are comparable to the short mountain ranges of the adjacent land. The submarine relief is also comparable in magnitude, but not in the intricacy of detail. The San Bernardino Range rises about 9000 feet above the adjacent basins, and the San Juan seamount rises about 10,000 feet above the adjacent ocean floor. Santa Cruz Island rises almost 9000 feet above the floor, and Catalina Island about 6000 feet.

Some of the relief features have flat tops. The most extensive are banks under about 3000 feet of water. Another group of flat-topped seamounts ranges in depth from 1200 to 3480 feet.

Continental Shelf

Shelf. North of Point Conception, the basin and range type of topography on the sea floor composes itself into a continental shelf generally not over 500 feet deep. Off central and northern California, the shelf is about 25 miles wide, and off Oregon and Washington, somewhat less. The borderland of southern California, after deepening southward, shoals again and abuts against the 80-mile-wide shelf of Sebastian Vizcaino Bay of central Baja California. From Sebastian Vizcaino Bay southward, a distinct shelf and straight shelf slope extend all the way to the southern tip of the peninsula. See map, Figs. 32.1 and 32.5.

The shelf zone continues fairly regularly along the coast of British Columbia and southeastern Alaska to a point off Yakutat Bay, where it turns southwestward along the Aleutian Islands and borders the Aleutian trench. It is a submerged surface of great glacial valleys off British Columbia and southeastern Alaska (see Fig. 17.18). Along the Aleutians, it is over 100 miles wide in places, and generally less than 500 feet deep.

Longitudinal depressions just off shore in the shelf of southeastern Alaska (off Yakutat Bay and Cross Sound) are interpreted to be due to faulting incident to the Pleistocene uplift of the adjacent ranges (Holtedahl, 1958).

Shelf Slope. From Yakutat Bay, Alaska, to Baja California, the shelf and basin and range borderland are terminated oceanward by a slope of great proportions. The decline where greatest extends from the brink at 500 feet to the base at 10,000 feet. In places it is sufficiently steep to be comparable with the Sierra Nevada scarp, and hence considered by

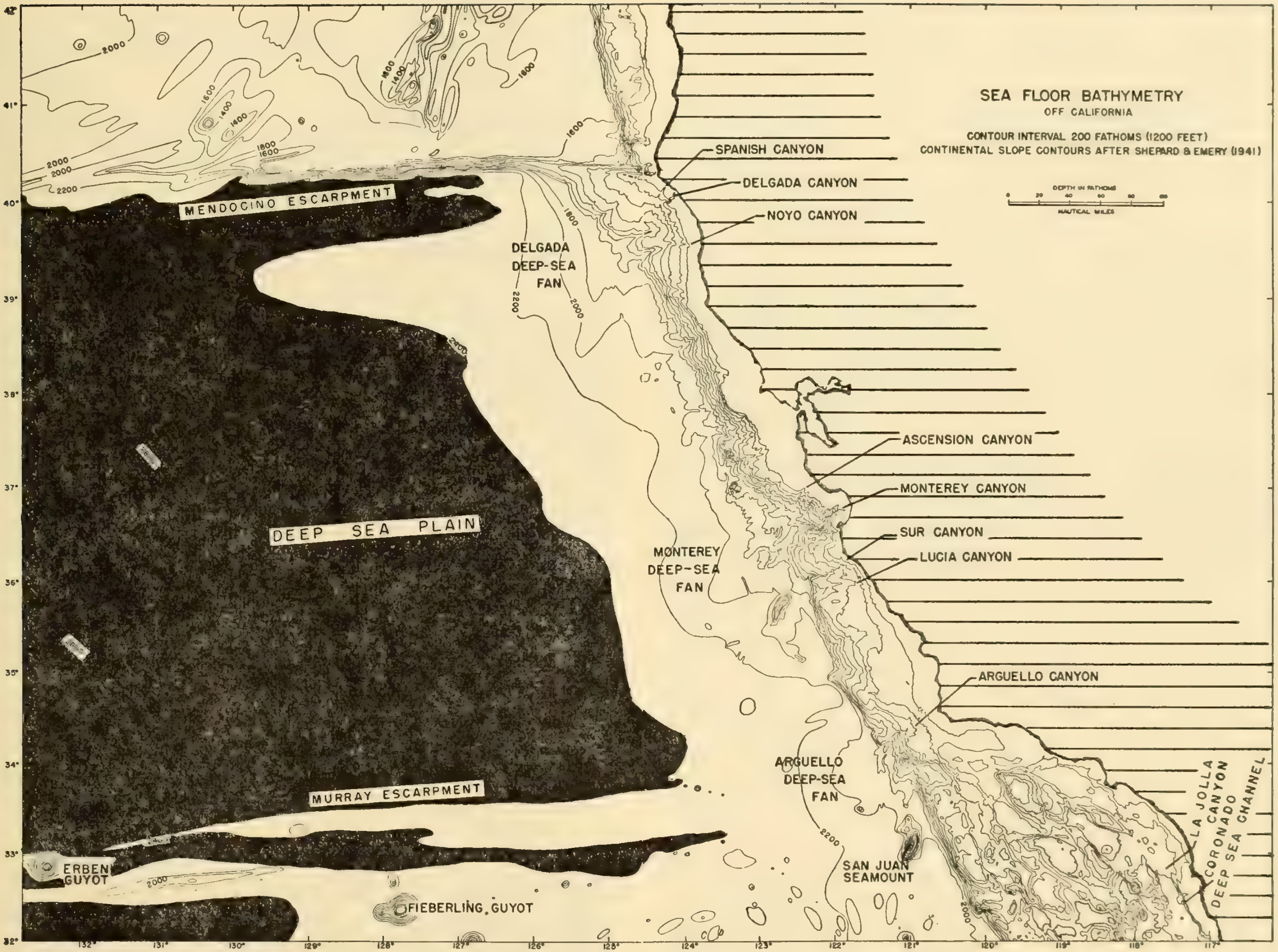


Fig. 32.2. Bathymetric chart of sea floor off California. Reproduced from H. W. Menard, 1955b.

Shepard (Shepard and Emery, 1941) to be a fault scarp. In other places it is not so steep and does not appear to be due to faulting.

One of the most fascinating discoveries of echo sounding is canyons that gash the shelf and its outer slope. Some of them are veritable gorges. A V shape is characteristic. There are about 66 of these submarine gorges or canyons along the California coast, and they are spaced irregularly at distances of 10 to over 50 miles. Most of the large canyons head within 3 to 5 miles of the present shore, but a few extend to within half a mile. Some of the smaller ones head 30 miles out. The longitudinal gradients are high and compare closely with stream gradients whose canyons have been cut in fault scarps. The gradients average about 4 degrees, are steeper near their heads, and gentler in the lower reaches and the longest canyons. The canyon bottoms are as continuous down hill as those of typical mountain canyons, at least out to depths of 6000 to 9000 feet, where the gentler outer slope may in places have suggestions of shallow basins.

The depth of the canyons is variable. The long Arguello Canyon west of the Santa Barbara basin starts in four tributaries, each only 300 feet deep. These shallow gorges trench the shelf slope out to where it is 3000 feet deep. Each of the tributaries is about 15 miles long. They converge into a single canyon which, in another 15 miles, is nearly 2000 feet deep. At about the 5000-foot depth contour the V widens, although the canyon is over 1000 feet deep at the point. The canyon turns southward, and may be followed down to 11,700 feet below sea level.

Another great submarine canyon, the Monterey, begins in tributaries in the Bay of Monterey which are 2000 feet deep a mile below their heads. The main canyon is 3000 to 4000 feet deep, and it trenches the shelf margin as a narrow V-shaped valley to a depth of 9000 feet, where it widens and shallows. It turns southward at this point and may be traced clearly still deeper to 11,000 feet below sea level.

Long stretches of the outer slope of the continental shelf are not dissected by submarine canyons. One stretch is north of Arguello Canyon between latitudes 34° and 35° 40′, and another is between Eel Canyon, off Cape Mendocino, and the Columbia River. Gentle slopes are in part characteristic of these margins, and Shepard points out that canyons are not so common on gentle offshore slopes as on steep ones.

The continental shelf north of the Aleutian trench, quoting from Murray (1945), is:

> . . . approximately defined by the 100-fathom contour. The maximum width of the shelf, 120 miles, is in the vicinity of Kodiak Island. To the northeast and southwest, the shelf narrows to a few miles as it converges with the major land features. The coast line is generally irregular and precipitous, although there are interspersed occasional areas of low relief. Only two principal rivers, the Susitna emptying into Cook Inlet and the Copper northwest of Cape St. Elias, discharge sediment onto the shelf or into the inland waters.

Deep-Sea Fans. Turbidity currents debouching from the mouths of submarine canyons have built large cone-shaped deposits called deep-sea fans. See Fig. 32.2. Their volume is usually many times the volume of material that could have been eroded from the canyons, so it is presumed that much sediment is contributed by shoreline processes to the heads of the submarine canyons (H. W. Menard, Jr., 1955), which then moves down the canyons to the fans below. The fans bury much or all of the previous relief on the deep-sea floor and produce smooth gentle slopes.

Origin of Submarine Canyons. The submarine canyons of the California shelf were postulated to be drowned subaerial valleys, smothered by sediment, and excavated by glacial and recent turbidity currents (Daly, 1936). Shepard (1952) contends that turbidity currents are not potent enough to erode the canyons and suggests that drowned river valleys have been kept permanently open by the turbidity currents during the process of submergence. Kuenen (1953) counters that this process does not explain all types of submarine canyons. Figure 32.3 is a reproduction of his conception of the different kinds of submarine canyons off the California coast, and he comments as follows about their origin:

> Instead of assuming that drowned valleys were perpetuated by sliding and turbidity currents, which have no ability to erode, it is suggested that the ancient land surface was first smothered; later the poorly consolidated covering materials were eroded during the Ice Age, and to some extent in postglacial times to form the submarine canyons.
>
> Some localities were particularly favorable to the generation of turbidity currents because of incompletely buried topographic depressions, local supply

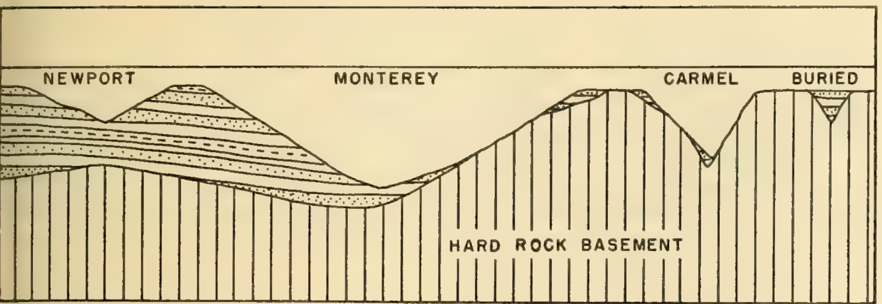


Fig. 32.3. Possible constitution of different submarine canyons off California. After Kuenen, 1953.

of sediment by rivers, and coastal configuration. Some narrow rocky land valleys were thus swept out (Carmel, Scripps, etc.), but the majority of old valleys may still lie buried in the terrace beneath sediments.

In some cases the turbidity currents only cleaned parts of the old valleys where these happened to offer small resistance. But other parts of these valleys did not conform to the requirement of following the present slope. Such parts remained buried.

Elsewhere a new valley cleaned off along its wall some small part of an ancient mountain slope, without conforming to the original drainage pattern. This may be the case for Monterey Canyon, which has granite overlain by sedimentary rock on one wall opposite a wall which has yielded only mud or soft sedimentary rock; or for Dume Canyon with basalt on the east side and mud with calcareous shale on the west.

Origin of the Continental Shelf Slope. The imposing slope has been ascribed to faulting, and the shelf itself primarily to wave cutting (Shepard, 1948). The Atlantic terrace, however, has been described as developed by sedimentation and isostatic subsidence caused initially by the sedimentary load (Kuenen, 1950). This theory of origin is amply attested locally, for instance, by the Mississippi delta building and consequent subsidence in the Gulf of Mexico. See Chapter 36.

We have to deal primarily with the consequences of orogeny in the marginal belts of the continent and then secondarily, with the processes of erosion, sedimentation, and epeirogeny in explaining the existing continental shelf and shelf slope. It is not clear yet what an orogeny such as the folding of the strata of the Coast Ranges of Oregon and Washington does to the continental shelf slope, or in what condition it is left, but in any consideration, the gradation from continental crust to oceanic crust will result isostatically in a surficial (submarine) slope toward the ocean. This may then be altered by erosional, depositional, and epeirogenic processes. In the previous discussions of submarine canyons and slope aprons or fans, and in subsequent discussions of the Aleutian and Middle America trenches and the possible faulting off Oregon and northern California the nature of the secondary processes is illustrated.

ALEUTIAN TRENCH

The Aleutian trench is a narrow depression in the ocean floor paralleling the convex side of the Kenai and Alaska peninsulas and the Aleutian volcanic island archipelago. See Figs. 32.1, 32.4, and 39.1. It extends from Yakutat Bay in the Gulf of Alaska westward to Attu Island, a distance of over 2200 statute miles. It has a maximum depth of 25,000 feet. According to Murray (1945):

> The vertical relationship between the crest of the conspicuous mountain features and the floor of the trench is shown in Fig. 32.4. An approximate difference of 28,000 feet exists throughout most of the region. The greatest single known difference throughout the entire arc exists slightly east of the mid-section and is centered at Unimak Island, where Shishaldin Volcano (9372 feet) rises 32,472 feet above the floor of the trench (about 110 miles distant). . . .

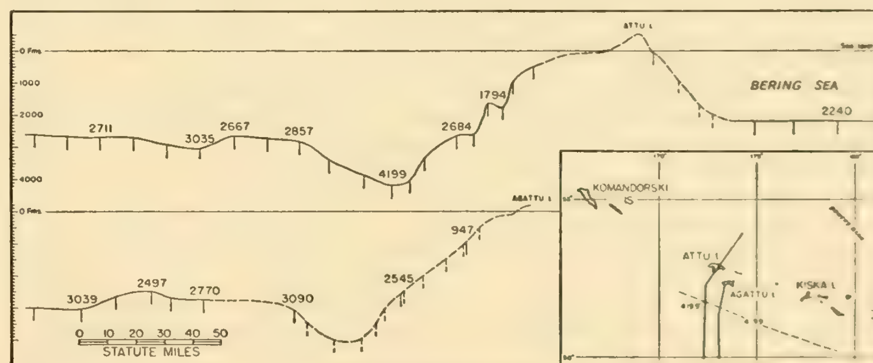


Fig. 32.4. Profiles of the Aleutian trench in the vicinity of Attu and Agattu Islands, western end of the Aleutians. After Murray, 1945.

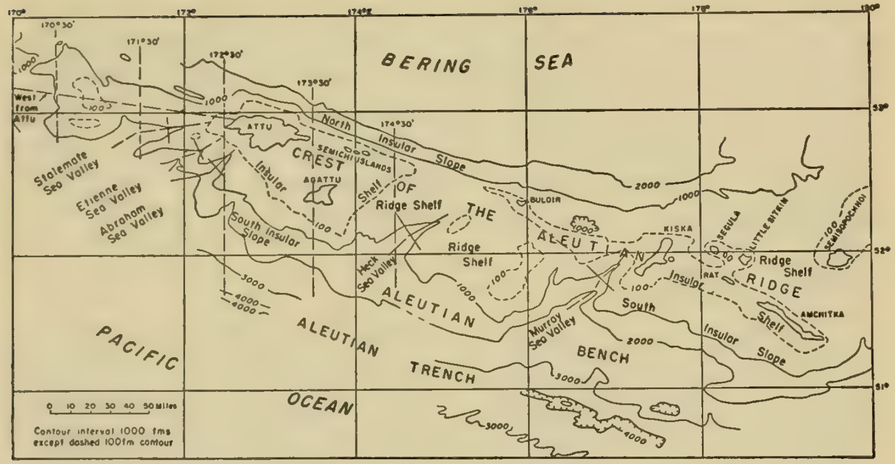


Fig. 32.5. Submarine contour map of west end of Aleutian Ridge. Reproduced from Gates and Gibson, 1956.

The continental slope comprising the inner north side of the foredeep is considered approximately as the area between the 100-fathom contour (50-fathom contour frequently applicable) and the floor of the trench. It ranges from 20 to 70 miles in width, is narrower near Cape St. Elias, and widest off Umnak and Unalaska islands. South of Unimak Island, a pronounced widening, herein termed the "Aleutian Bench," exists between the 2000- and 2500-fathom contours and extends westward to Umnak Island. This bench is approximately 20 miles wide and 170 miles long. The bench lies several hundred fathoms higher than the top of the outer seaward side of the trench.

The average slope of the north face or continental slope is 3°–4° and terminates in depths ranging from around 2,000 to 4,000 fathoms. Steeper slopes, however, are found in limited areas or between successive soundings. When the slope exceeds 30°, it usually occurs near the bottom of the trench where the profiles show an abrupt slope or escarpment as, for instance, the apparent escarpment off Cape St. Elias.

The surveyed slopes on the north and south sides of the Umnak Island locality differ materially with respect to relief and rate of descent. The north side of the island is characterized by long valleys and ridges in the deeper area. For instance, the maximum seaward distance of the 1000-fathom curve on the north side of Umnak Island is 45 miles, whereas that on the south is barely 5 miles.

The floor of the trench, 20 to 70 miles off the edge of the continental shelf, undulates, but steadily descends in the 1000-mile stretch from Cape St. Elias to Umnak Island. In many profiles, the converging side slopes of the trench meet in a narrow area defined by one or two soundings at, or close to, the base of the continental slope.

The gentle incline of the trench terminates at about 2,000 fathoms, off Cape St. Elias. The trench, however, continues eastward across the continental slope and then, apparently, is continuous with a depression extending across the continental shelf toward Yakutat Bay. The delineation of the 100-fathom curve on the shelf here is inconclusive, as it is controlled by only a few widely spaced soundings. A bar with depths of 8 to 16 fathoms extends entirely across the entrance to Yakutat Bay. Depths as great as 167 fathoms, however, are found about 4½ miles inside the bay.

Detailed contouring of the west end of the Aleutian Ridge has led Gates and Gibson (1956) to postulate that the submarine topography reflects the structure. The Aleutian Ridge with its islands is shown in Fig. 32.5, and the suggested structure in Fig. 32.6. The geology of the Aleutian Islands will be discussed in Chapter 39, but suffice it to say here that Attu, Agattu, the Semichi Islands and the southern part of Kiska lack young stratovolcanoes and are composed of pre-middle Tertiary rocks and subordinate amounts of upper Tertiary coarse clastic sediments and subaerial lava flows. They owe their height to faulting and alpine character to vigorous erosion. The fault pattern of Attu and Agattu, particularly, is obvious and intricate. It has led to the interpretation of submarine features as fault reflections.

Four principal topographic provinces are recognized: (1) The Crest of the Aleutian Ridge contains the Aleutian Islands, the Insular Shelf at depth ranging from present shore lines to 70 fathoms, and the Ridge Shelf at a depth of 100 to 500 fathoms, all apparently the result of subaerial and marine erosion since the middle Tertiary and of glaciation in the late Pleistocene. (2) The Insular Slopes form the sides of the Aleutian Ridge. The North Insular Slope is a long, steep, linear scarp that probably marks a major fracture in the earth's crust. The South Insular Slope appears to be a broad, faulted and warped arch containing numerous steep-sided linear sea valleys and canyons. Many of these traverse the south slope at an angle to the maximum regional gradient, and several line up with observed faults on the island. These linear topographic features probably mark fault zones. (3) The Aleutian Bench is a prominent step in the general slope from the islands to the Aleutian Trench, and its inside edge may be the trace of a thrust fault. (4) The arcuate Aleutian Trench has a steep north side, a flat floor at a depth of about 4000 fathoms, and a south side containing an *en echelon* topographic pattern. The Trench perhaps marks a major thrust zone dipping north beneath the Aleutian Ridge.

A structural interpretation of the submarine topography suggests that the

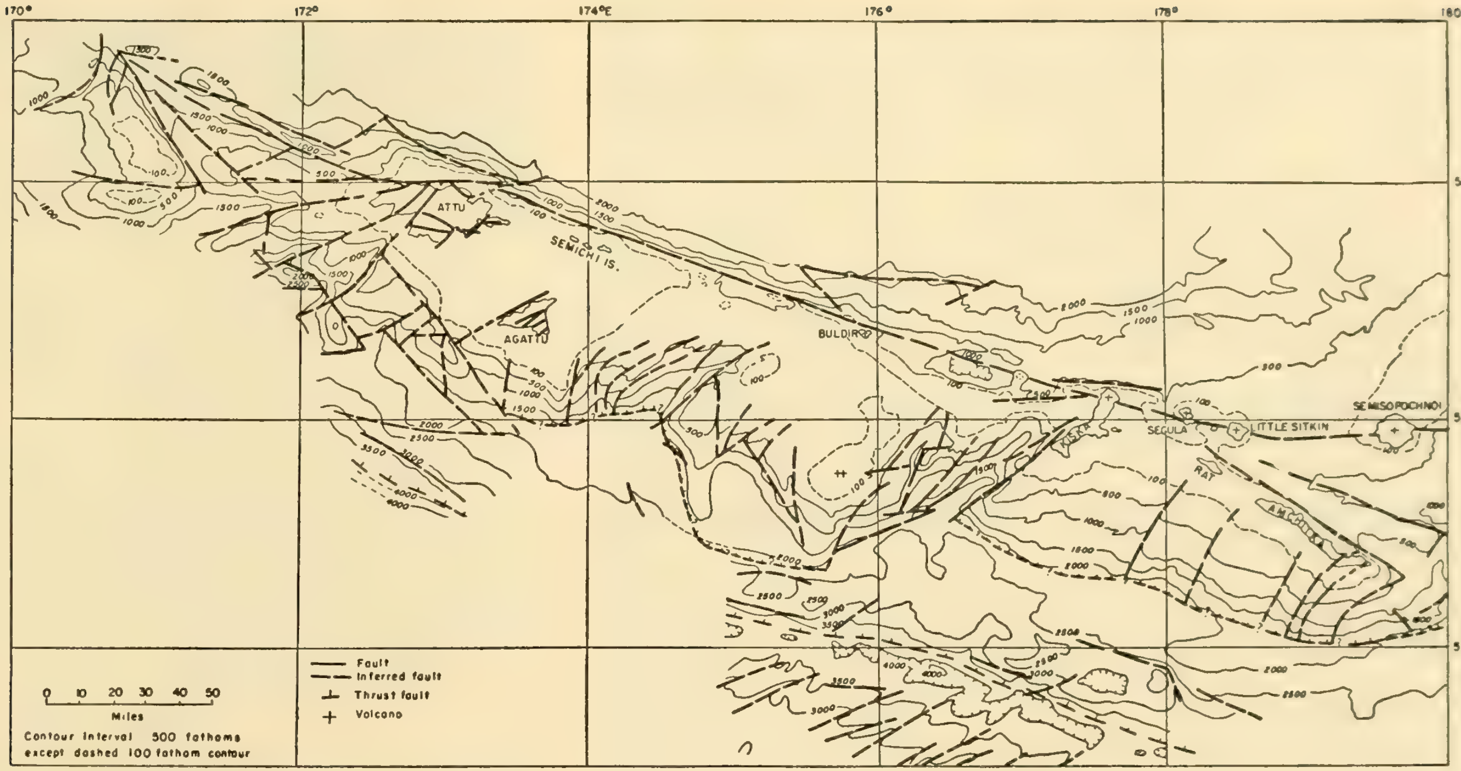


Fig. 32.6. Postulated faults of end of Aleutian Ridge and trench. Reproduced from Gates and Gibson, 1956.

western part of the Aleutian Ridge is an arched and faulted asymmetrical wedge bounded by a northward-dipping normal fault on the north and by a northward-dipping zone of reverse faults on the south. Formation of this wedge probably began with major uplift and faulting of the western Aleutian area during the middle Tertiary, and the many earthquakes and active volcanoes in the Aleutian arc today indicate that deformation is still continuing (Gate and Gibson, 1956).

The structure of the ridge as Gates and Gibson speculate is shown in Fig. 32.7.

BERING SEA FLOOR

The Bering Sea is a closed triangular-shaped basin bounded by two continents and the arc of the Aleutians. About half the area is continental shelf, and half lies at depths of 1600 to 2240 fathoms. The greater depths are in the southwestern portion. The maximum depth recorded, 2240 fathoms, lies 45 miles northeast of Attu Island, and is approximately 2 miles above the floor of the trench on the south side of the Aleutian Islands. See Figs. 32.1 and 39.10.

The deep division of the Bering Sea is marked by a submarine range that takes off northward from the Aleutian arc and veers westward. It is 300 nautical miles in length, 60 miles in width, and rises in one place 12,156 feet above the bottom. It is known as the Bowers Bank Range and supports Semisopochnoi Island and the Petrel Bank, as well as Bowers Bank.

The Pribilof Islands emerge from the shelf of the Bering Sea, which

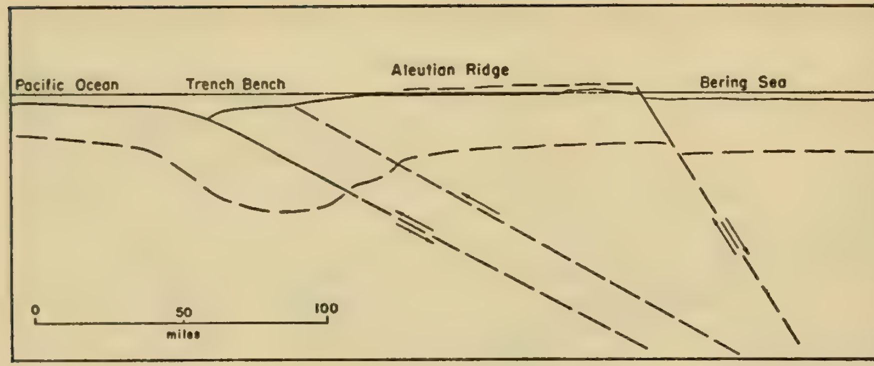


Fig. 32.7. Speculative and diagrammatic cross section of western end of the Aleutian Ridge and Aleutian trench. Reproduced from Gates and Gibson, 1956.

in large measure appears to be the great delta of the Yukon and Kuskokwim rivers. See Chapter 39.

PACIFIC FLOOR OFF MEXICO AND CENTRAL AMERICA

Middle America Trench

The Middle America trench is continuous at depths greater than 14,400 feet for 1260 miles, except for two submarine volcanoes which lie in the trench. (See Figs. 32.8 to 32.10). Northwest of Acapulco the trench is generally U-shaped in cross section, with a steeper shoreward flank and a flat bottom suggesting sedimentary fill. Off Guatemala for a distance of 380 miles it is over 18,000 feet deep with a maximum sounding of 21,000 feet. Thence southeastward it shoals gradually to merge into the sea floor off Costa Rica. The southeast segment is also asymmetrical in cross section, but V-shaped with irregular bottom, in contrast to the flat bottom northwest of Acapulco.

Along the trench as explored to date, a series of breaks in slope or terraces suggests a downwarped or downfaulted shelf below the more normal shallow shelf. Faulting across the shelf may have been important south of the Isthmus of Tehuantepec (Fisher and Shor, 1959).

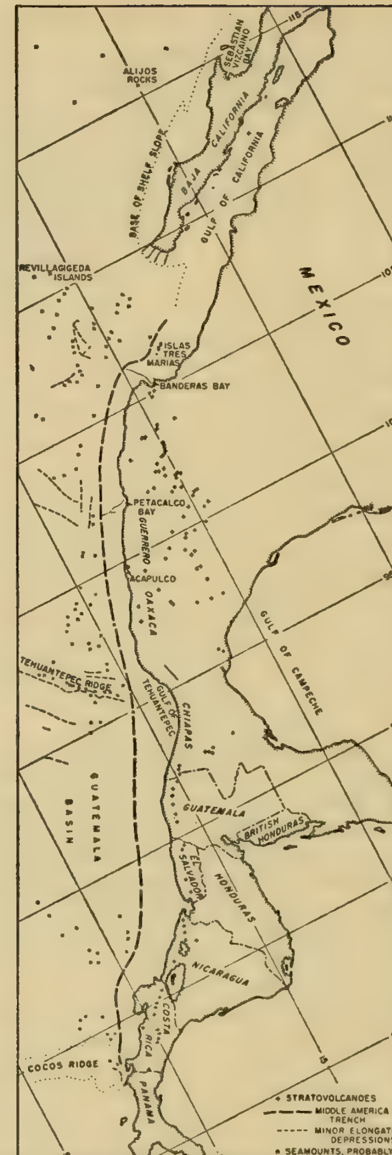


Fig. 32.8. Middle American trench and related features. Compiled from Fisher 1961, and Shor and Fisher, 1961. Rows of dots are submarine canyons.

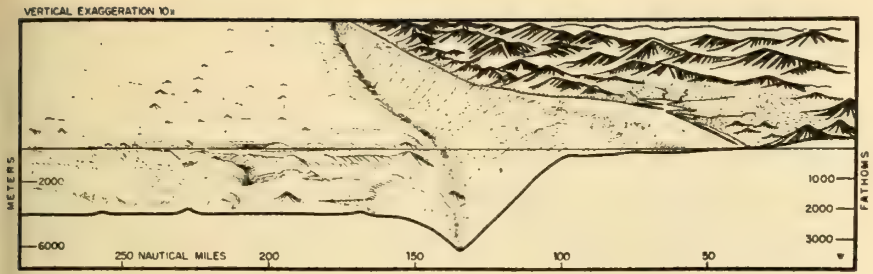


Fig. 32.9. View of Middle American trench to northwest from Gulf of Tehuantepec. Tehunantepec Ridge is in left foreground. Reproduced from Fisher 1961.

Tehuantepec Ridge

A northeast-southwest trending band of ridge and trough topography, 60 miles wide, separates the 10,800–11,400-foot sea floor outside the trench off southern Mexico from the 12,600–13,200-foot Guatemala basin. This zone has been traced from several hundred miles offshore to an intersection with the trench near the west side of the Gulf of Tehuantepec, and has been called the Tehuantepec Ridge (Figs. 32.8 and 32.9).

Ocean Floor and Seamounts

The ocean floor outside the trench is fairly flat except for numerous seamounts which undoubtedly are volcanic cones. The map of Fig. 32.8 shows the distribution of the seamounts charted by Fisher and Shor (1959) and also the volcanic cones of Recent or Pleistocene age on land in southern Mexico and Central America as far as the writer has been able to locate them from the literature.

The Guatemala basin, which is about 1800 feet deeper than the floor north of the Tehuantepec Ridge, shoals to the southeast. It contains few volcanoes whereas a row of majestic active and dormant volcanoes lies opposite on land and stretches from southern Chiapas across Guatemala, El Salvador, Nicaragua, and Costa Rica. Volcanism in Mexico is discussed in Chapter 35.

As far as known the distribution of volcanoes on the ocean floor southwest of the trench is random, although one or two rows seem apparent. None of the seamounts has been recognized as beveled, so it is not possible to infer vertical movements of the ocean floor such as in the Mid-Pacific Mountains, described on following pages.

Crustal Structure

Three seismic refraction stations were taken along the axis of the trench west of Acapulco and two along its axis off Guatemala and El Salvador. Another station was shot on the shelf and one 60 miles seaward of the trench off Guatemala. Upper mantle velocities appear on all lines (Fisher and Shor, 1959).

Thick sediments were found in the Tres Marias basin off Manzanillo and at the shelf station off Guatemala. On a section normal to the trench off Guatemala, the depth below sea level to the Mohorovicic discontinuity in the trench zone is 16 kilometers, and in the shelf area 17 kilometers. Below the sea floor the crust thickens from 5 to 7 to 10 to 17 kilometers along this section (Fig. 32.11).

The Mohorovicic discontinuity is deeper and the crust below the sediments thicker under the two southern stations than under the two central trench stations. The mantle is deeper under the Tres Marias basin, where thick sediments (1½ kilometers) are found, than under the central stations.

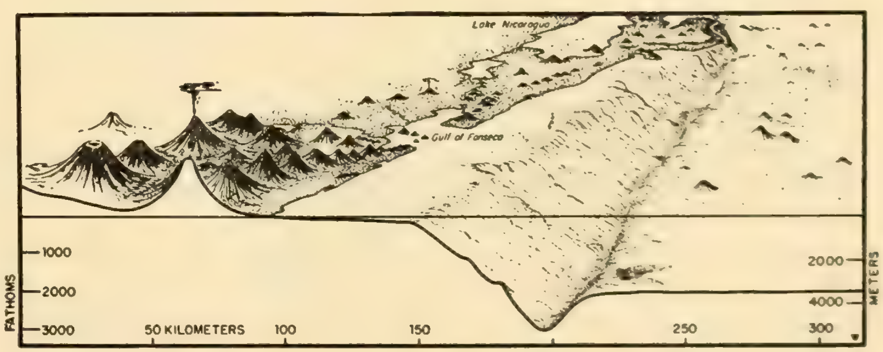


Fig. 32.10. View of southeastern end of Middle America trench. Reproduced from Fisher 1961.

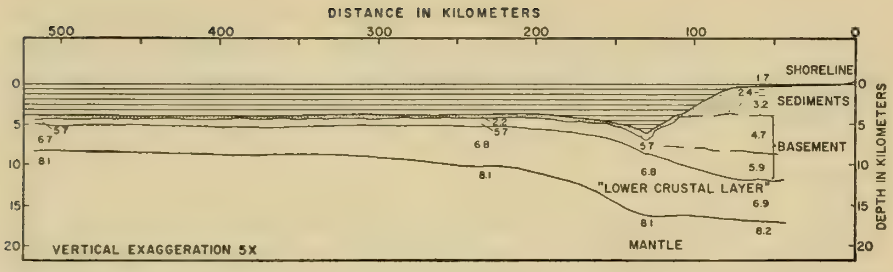


Fig. 32.11. Crustal layers across Middle America trench after Shor and Fisher, 1961. Numbers represent wave velocities in kilometers per second.

Age of Trench

The Gulf of Tehuantepec marks a major change in trench configuration and possibly in age. Northwest of Tehuantepec the flat trench bottom suggests a greater age than the deep V-shaped profile southeast of the Gulf. Thicker crustal layers and a bordering volcanically active coast also mark the younger division. The zone of ridge-and-trough topography, the Tehuantepec Ridge, trending southwest from the point of change may be another evidence of the division of the trench into older and younger parts.

FRACTURE ZONES

Four great bands of linear relief features, named fracture zones (H. W. Menard, 1955), have been discovered in the northeastern Pacific basin. They are the Mendocino, Murray, Clarion, and Clipperton, and are shown on the map of Fig. 32.12. A lesser zone, the Pioneer Ridge, is labeled on Fig. 32.16. It had not been surveyed well at the time the map of 32.2 was constructed.

The zones range from 1400 to 3300 miles long and average 60 miles wide. The Mendocino and Murray stretch across the Pacific floor to the Hawaiian Ridge. They follow great circle courses and are approximately parallel. Topographic relief within the fracture zones is characterized by large seamounts, deep narrow troughs, asymmetrical ridges, and escarpments. Two escarpments are about 1 mile high and more than 1000 miles long. See Fig. 32.2.

The Clipperton fracture zone is more varied and irregular than those to the north (Menard and Fisher, 1958). The western half consists of narrow ridges and low seamounts, but the eastern is dominated by an enormous ridge, about 60 miles wide, 330 miles long, and 8000 to 10,000 feet high. A trough about 10 miles wide and a mile deeper than the surrounding region borders the ridge. See Fig. 32.13.

The over-all easterly trend of the ridge is complicated by a southeasterly cross trend indicated by the alignment of volcanoes, by orientation of minor ridges on the south side of the main ridge, and by the marked change in trend of the main ridge at its eastern end. Clipperton Island, the only feature in the whole Clipperton fracture zone that reaches the sea surface, is one volcano

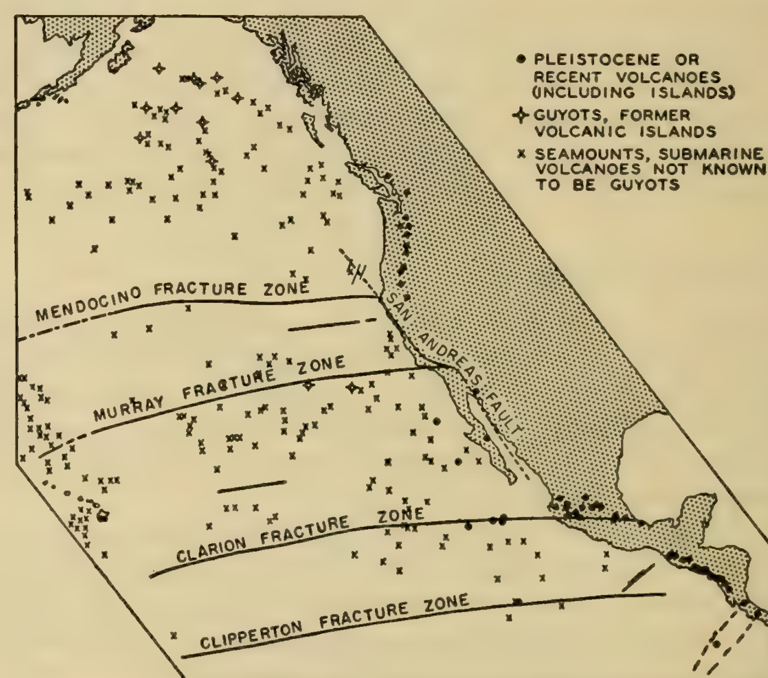


Fig. 32.12. Fracture zones and seamounts of northeastern Pacific. Reproduced from H. W. Menard, 1955b. Also volcanoes of adjacent coastland.

on a cross trend. The maximum relief of the Clipperton Ridge is 18,000 feet from Clipperton Island to the deepest spot in the trough at 2,960 fathoms.

DEEP SEA PROVINCES

Gulf of Alaska Seamount Province

The northernmost division of the northeastern Pacific basin is the Gulf of Alaska Seamount Province (Menard and Dietz, 1951). Its northwestern boundary is the Aleutian trench and its western the continental shelf slope, which here is only about 8000 feet high. A rather steep apron flattens seaward and appears to be a graded profile. The apron and smooth deep-sea floor are interrupted by thirty-six majestic submarine volcanoes. Eleven of these are guyots, and their flat tops indicate they were once truncated by erosion. Most of them are now about 2500 feet below sea level and some are much deeper, so it is concluded that a like amount of subsidence has occurred since the truncation.

The region is seismically inactive, and the topography is old with a thick apron of sediment evidently across the entire province. Major subsidence of the region is postulated but some time in the geologic past, possibly Cretaceous.

Ridge and Trough Province

The continental slope of the Ridge and Trough Province is about 1½ miles high and is dissected by several well-known submarine canyons. An apron of sediment spreads from the base of the slope off Queen Charlotte Island in the northern part of the province, but a long, narrow, seismically active trough lies between the apron and the base of the slope. Evidently the top of the apron has been faulted down so recently that sediment moving out from the continent has not yet filled the trough to re-establish an even gradient seaward (H. W. Menard, 1955).

The sea floor presumably was block-faulted into long thin ridges which trend northeast or north. From the ridges rise a few submarine volcanoes some of which are only a few fathoms below the surface, but most crossings of the ridges indicate steep-sided, low blocks, unlike volcanoes.

The long ridges roughly parallel the continental slope and guide the flow of turbidity currents moving sediment out from the continent. One of several leveed channels on the otherwise smooth plain at the base of the continental slope off Oregon was traced southward for almost 200 miles. Apparently the turbidity currents cannot surmount the ridges to flow west (direction of the regional slope) but are diverted southward to a divide through which they again flow westward or fan out to fill low spots on the downstream side of the ridges. A few basins appear entirely ringed by high ridges so that turbidity flows moving along the bottom cannot fill them with sediment. These basins are thousands of feet below the level of the surrounding alluvial plains formed by deposition from turbidity currents; their bottoms are irregular, which suggests that deposition from suspension in the main mass of the ocean may be much slower than deposition from turbidity currents moving in concentrated clouds along the bottom.

Deep Plain

South of the Mendocino escarpment the sea floor is about half a mile deeper than it is to the north, and it is called the Deep Plain. It is bounded on the south by the Murray escarpment, and south of the Murray escarpment the sea floor is roughly a quarter of a mile higher than it is to the north.

The continental slope off central California forms the eastern boundary of the Deep Plain. It drops off abruptly to a depth of more than 2 miles, and three great deep-sea fans form an apron which grades imperceptibly into the gently sloping Deep Plain at a depth of about 2½ miles. Crossing the fans are leveed and unleveed channels.

The Deep Plain is unique in that it appears to contain few seamounts. Five seamounts, probably volcanoes, rise from the continental slope

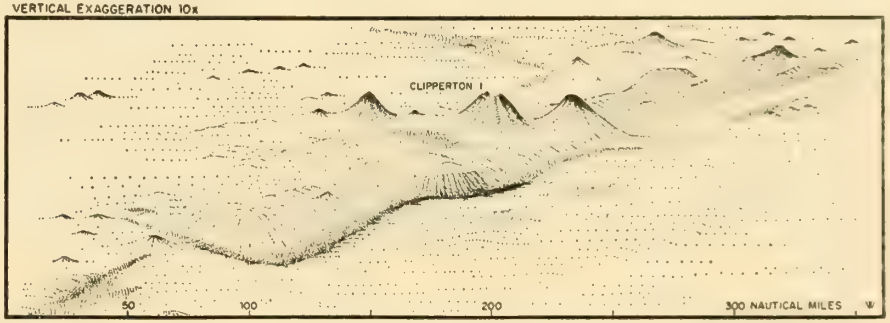


Fig. 32.13. View to southwest toward Clipperton Island and the Clipperton Ridge. Reproduced from Menard and Fisher, 1958.

bordering the area, but they trend parallel to the coast and may be genetically unrelated to the deep-sea floor. (H. W. Menard, 1955).

Baja California Seamount Province

South of the Murray fracture zone a mountainous area, studded with volcanoes, forms the Baja California Seamount province.

The continental slope drops off abruptly to a depth of about 2 miles. It is irregular but does not appear deeply dissected by canyons. A smooth apron a few tens of miles wide lies at the base of the slope in some places. Off the southern half of Baja California the continental slope drops abruptly for 2–2½ miles into a series of long thin troughs a few hundred fathoms below the general level of the deep-sea floor to the west. The troughs are flat-bottomed indicating a fill of sediment.

Widespread vulcanism, particularly recent vulcanism, characterizes the province. Guadalupe Island comprises a group of eroded Late Tertiary or Quaternary volcanoes. Alijos Rocks are three steep-sided remnants of a large volcanic cone. Volcanic islands are so rare in the northeastern Pacific basin that these deserve special consideration, but the evidence supporting unusual vulcanism comes chiefly from submarine volcanoes. Of 51 seamounts, 15 are more than 1 mile high, and every expedition crossing the province finds new seamounts. Seven seamounts have been surveyed, and Jasper and Henderson have been dredged. The volcanoes are typical isolated cones with steep sides and pointed tops. None are guyots with wide flat tops. Henderson Seamount appears to have a flat top at 220 fathoms, but the area is only half a square mile, and this is too small to demonstrate that a sharp peak has been planed off. However, hundreds of pounds of coarse, basaltic gravel were dredged from the top of this seamount, and a large fraction of subrounded and subangular pebbles and cobbles suggests wear in the surf zone.

Contrasting strongly with the smooth floor of the Deep Plain to the north, the Baja California Seamount province is irregular. Recorded echo soundings show thousands of miles of jagged bottom in which the irregularities have a relief of 100–200 fathoms. The relief must be tectonic, but it is uncertain whether it is caused by vulcanism or faulting. The lack of a smooth blanket of sediment suggests either that the topography was formed relatively recently or that the rate of sedimentation is unusually slow. No large rivers carry sediment from southern California and Baja California into the ocean, and even the limited amount introduced by intermittent small rivers is trapped in the basins of the continental borderland or in the troughs off Baja California (Menard, 1955).

Constitution of Deep-Sea Crust

A seismic refraction survey by Raitt (1956) indicated that at a position in the Baja California Deep-Sea Province due east of Sebastian Vizcaino Bay (Lat. 27°24′N, Long. 121°35′W) in a depth of 4176 meters of water, the crust had the following velocity layers:

Thickness, km	Velocity, km/sec	
0.26	2.15	(Sediments)
0.93	5.88 ± 0.23	(Volcanics?)
6.24	6.96 ± 0.68	(Crust, gabbroic?)
	8.41 ± 0.43	(Mantle)

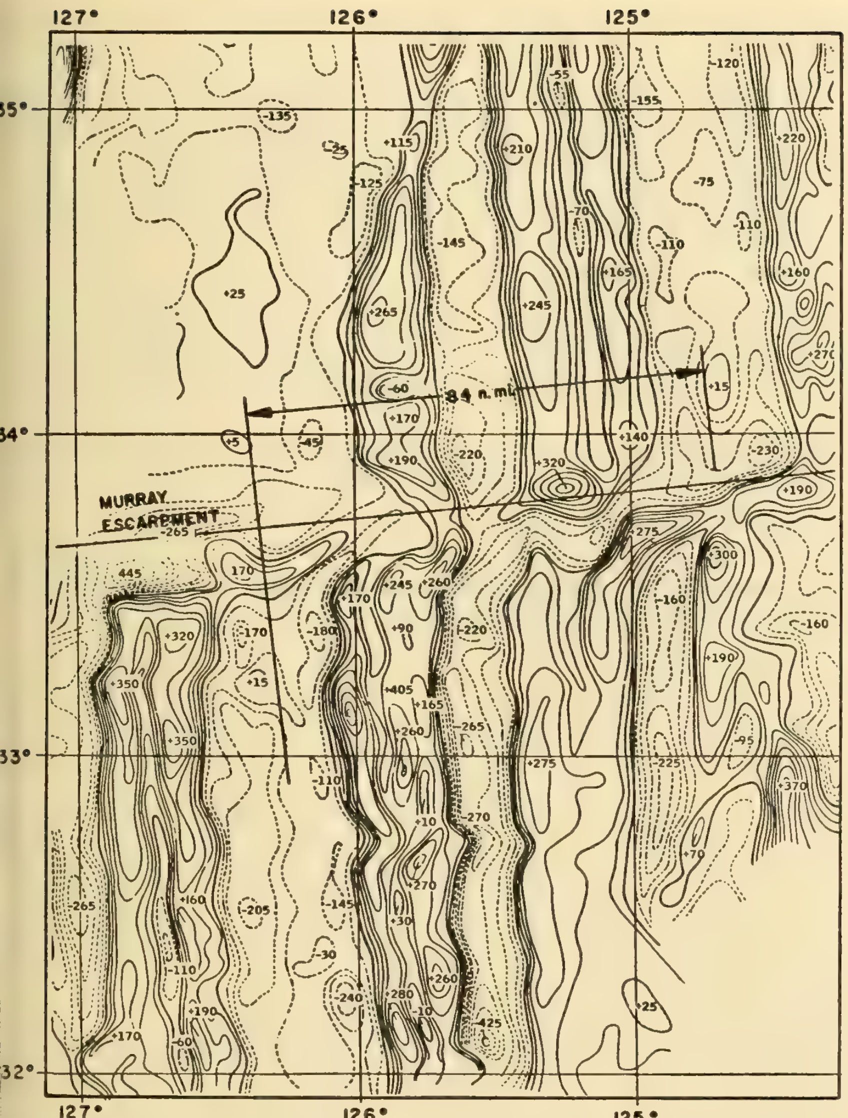
Mason uses similar figures in his analysis of magnetic profiles of the Deep–Sea Plain. See subsequent pages and Fig. 32.17.

Magnetic Intensity Surveys

Magnetic intensity surveys and contour maps have now been made of a large region off the western United States including a portion of the Deep Plain province and the Murray and Mendocino fracture zones (personal communication, H. W. Menard). The results are striking and tectonically significant.

Figure 32.14 is a sample of the magnetic intensity map and shows an area 350–400 miles out from the shore along the Murray fracture zone. The lines of equal magnetic intensity have been so adjusted that they do not reflect the increase of the earth's magnetic field across the area. The intensity highs and lows are in sharp zones about 15–25 miles wide and extend conspicuously and rather regularly in a north-south direction. This pattern is dominant west of a less intense and more irregular near-shore zone with a fabric to the north-northeast. Some of the strong north-south magnetic features have been contoured for a length of 370 miles on the Deep Plain (Menard and Vacquier, 1958; Mason, 1958).

Figure 32.15 shows the topography of the ocean bottom of the same area as Fig. 32.14. It will be seen that the Murray fracture zone is fairly narrow here and is reflected clearly in the magnetic intensity contours. It may also be detected that the zone is one of horizontal offset of the intensity pattern. This is brought out forcefully if an east-west profile curve of the anomalies field is plotted both north of the fracture zone and south of it. If the two profiles are then moved east or west they match well but in only one position. This is taken to mean that the



g. 32.14. Total magnetic intensity of an area off the California coast. Contour interval is gammas. Reproduced from Menard and Vacquier, 1959.

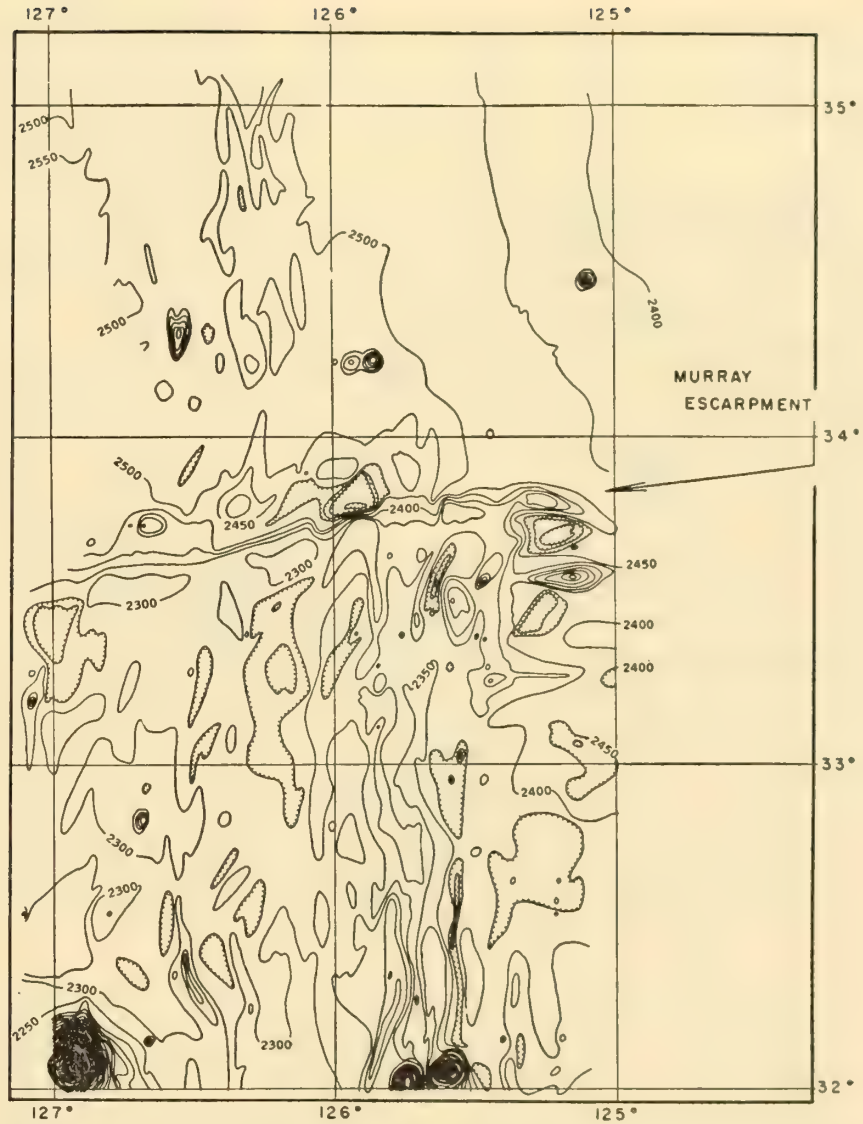


Fig. 32.15. Generalized topography of ocean bottom of Fig. 32.14. Reproduced from Menard and Vacquier, 1959. Contours in fathoms.

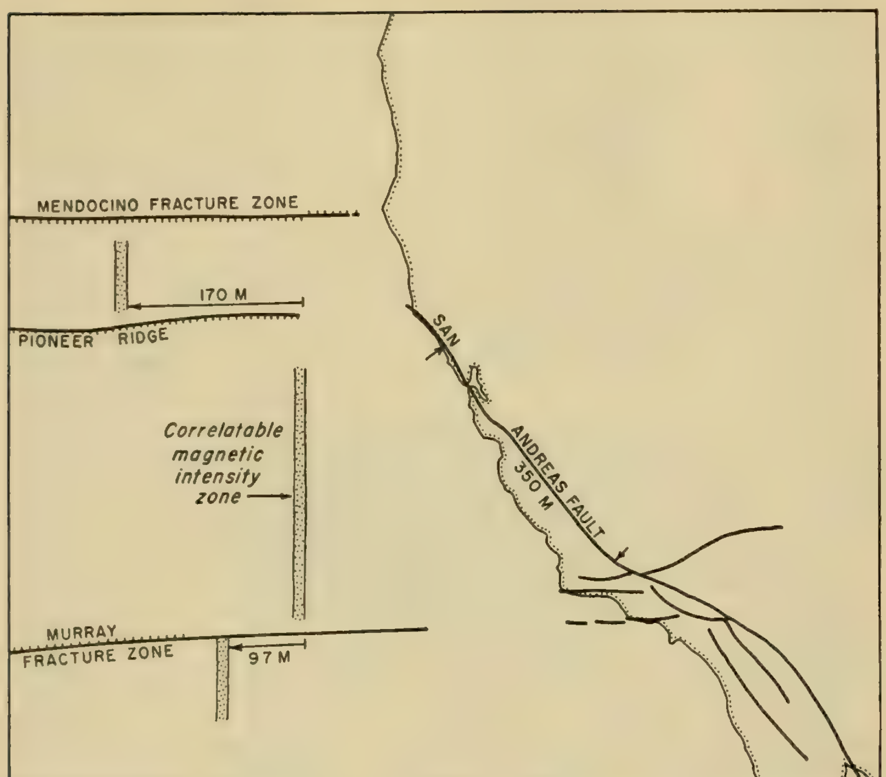


Fig. 32.16. Horizontal displacements along fracture zones indicated by the offset magnetic intensity field. Horizontal displacement along San Andreas fault also shown. After Menard (private map). Murray fracture zone offset by Mason (1958) and Pioneer Ridge offset by Vacquier, letter to *Nature*, 1959. Distances are in miles.

block of oceanic crust south of the Murray fracture zone has moved 97 statute miles westward. Likewise, the intensity pattern is offset along the Pioneer Ridge 170 statute miles (see Fig. 32.16) with the north block having moved west (Menard and Vacquier, 1958). The north block of the Mendocino fracture zone has moved the astonishing distance westward of 1250 kilometers, according to Vacquier *et al.* (1961). These considerable horizontal displacements are immediately thought of in connection with postulated strike-slip movement of the San Andreas fault, and the relation of the several postulated movements is shown in Fig. 32.12.

The magnetic expressions of the volcanoes are puzzling. Most all yield positive magnetic impressions in the intensity contours, but in no way are they as striking as the relief contours of the volcanic cones would suggest. Compare Figs. 32.14 and 32.15. They deflect the intensity contours of the dominant linear features or are superposed on them but are not sufficiently strong to make much of an impression. The magnetic effect is also variable according to Menard and Vacquier, who propose the variability to be due to the fact that some cones are built of fragmental material of lower intensity and some of massive flows of higher intensity.

The topography of the ocean floor has an irregular north-south fabric but it is of low relief and in striking contrast to the relief of the volcanic cones; yet its intensity contours are sharp and strong.

Regarding the cause of the anomalies it is evident that the distribution of rocks with different magnetic intensities must match the intensity pattern, with allowance made for depth and several magnetic factors.

In analyzing the profiles across the linear magnetic positive features the seismic refraction data of the area were first considered (Mason, 1958). The velocities and interpreted rock layers are shown in Fig. 32.17. The magnetic values are concluded to be compatible with those of basic igneous rock, which is here characterized by a high susceptibility and also a high intensity of remanent magnetization. As a consequence the "volcanics" layer is taken as the most likely seat of the anomalies, and calculations made to determine the depth, thickness, and lateral extent of basic igneous rock masses to produce the observed profiles. If the tabular mass is flat-bottomed, it would appear as in B, Fig. 32.17; if flat-topped, as in C, to produce the anomaly shown in A.

But how can we manage on a sound geologic basis the elongate tablets of basalt, diabase, or gabbro of the required shape and magnitude properly spaced and in parallel arrangement? The structure must be compatible with the subdued relief of the ocean floor. It should be pointed out that the topography of the northeast part of the area of Figs. 32.14 and 32.15 is particularly smooth and appears to be a graded alluvial

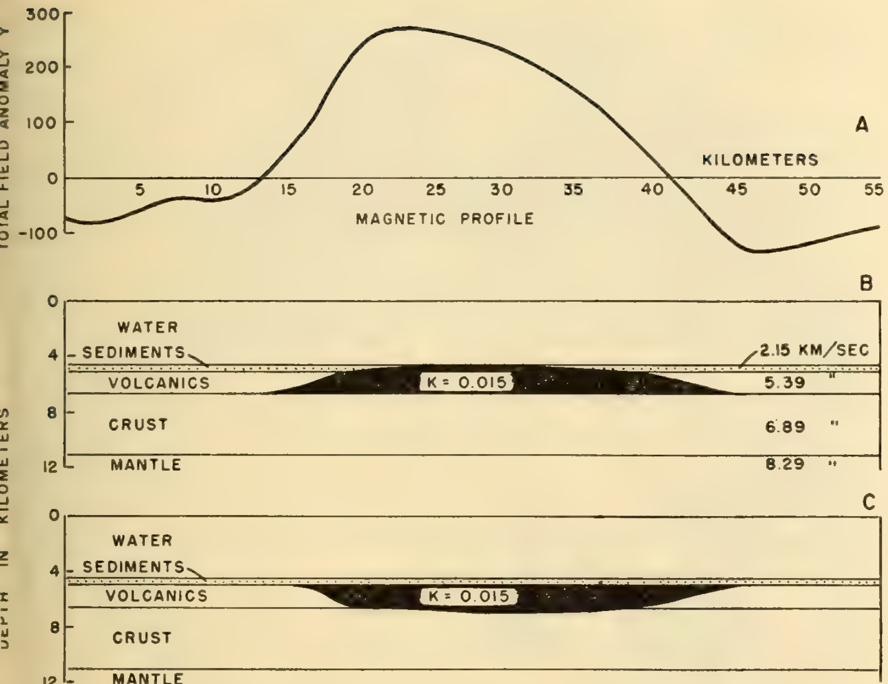


Fig. 32.17. Interpretation of magnetic profile (A), with flat base of basic igneous rock at 6.3 km depth (B), and with flat top at 5 km depth (C). After Mason, 1958.

profile or continental slope apron. If such, sediment has just about buried all previous existing relief there.

Menard (1955) thinks that the displacement along the fracture zone took place during Cretaceous or Tertiary time and that the structures causing the magnetic anomalies are older than the fracture zones. Possibly, therefore, the east-west fracture zones and the north-south structures are not related mechanically. Menard, Vacquier, and Mason suggest that parallel valleys were filled or partially filled with basalt and that later sediments were carried out by turbidity currents and by being spread in the remaining depressions still further reduced the relief. The cause of the parallel valleys and the nature of the eruptions is not considered, nor the relation to the other volcanic (?) rocks of the "volcanics" layer. More intensive seismic surveys will undoubtedly help in solving the problem.

HAWAIIAN RIDGE

The Hawaiian Islands are peaks of a ridge or swell built by volcanic action on the ocean floor. It has a relief from deepest ocean floor to top of peaks of nearly 32,000 feet, is about 150 miles across in the widest part

Fig. 32.18. Generalized topography around southern end of Hawaiian Ridge showing deep and arch, after Hamilton, 1957. Contours in fathoms.

and trends to the northwest (E. L. Hamilton, 1957). The islands are believed to have formed during the Tertiary with volcanic activity progressing southeastward. Present volcanic activity is confined to the island of Hawaii, which may have had its inception as late as the Pliocene (Stearns and Macdonald, 1946).

Submarine contouring has indicated a sag, the Hawaiian deep, adjacent to the ridge, which in its deepest part is about 3600 feet below an outer gentle arch. See Fig. 32.18. The bottom of the deep is above the level of the ocean floor beyond the arch.

The peripheral deep and arch are believed by Hamilton to be due to the loading of the earth's crust by the volcanic piles, and to consequent downbowing and lateral bulging.

MID-PACIFIC MOUNTAINS

A submerged relief feature known as the Mid-Pacific Mountains, extends southwesterly from Hawaii. There a series of flat-topped volcanic peaks, called guyots, are submerged 4200 to 5400 feet. The study of dredged samples from the flat tops yielding Upper Cretaceous, Paleocene, and Eocene foraminifera indicate that in Cretaceous time the guyots were a chain of basaltic islands, wave-decapitated with coral-rudistid reefs lodged on and among the erosional debris. Submergence followed to the depths indicated (E. L. Hamilton, 1956). The recognition of broad subsidence of the ocean floor in the magnitude of one mile is very significant in understanding the processes of mountain building.

Fig. 32.19. Circum-Pacific tectonics. Reproduced from Benioff, 1957.

CIRCUM-PACIFIC TECTONICS

In Chapter 29 the San Andreas fault and associated structures were depicted, and there the theories of Hill and Dibblee and of Benioff on the mechanics of formation were outlined. It is recognized that the major movement on the San Andreas fault has been right strike-slip movement. Hill and Dibblee (1953) have suggested a horizontal displacement of 560 kilometers.

Incident to the study of aftershock sequences Benioff (1957) recognized that the extent of faulting for earthquakes where the fault is not visible could be determined. Since the direction of slip can also be determined, a study of Circum-Pacific earthquakes leads to the presumed discovery that around the entire margin the slip is dextral as indicated in Fig. 32.18. Only for Antarctica are observations wanting.

Critical of Benioff's hypothesis of counterclockwise rotation of the Pacific basin crust, Chingchang (1958) points out that the section be-

tween Japan and the equator is rotating clockwise. The evidence lies in the study of several great earthquakes in the region and in the geology of known faults in the Philippines and Japan.

Whether or not the entire Pacific is moving counterclockwise, the problem arises along the North American margin: What is the relation of the dextral Pacific movement to the fracture zones? In a personal communication on the subject Dr. Benioff comments as follows:

I assume as a working hypothesis that the radial movements at the continental margins are expressions of growth of the continents by accretion of material from below by unknown processes. As the continents rise, the margins are driven over the adjacent oceanic masses by gravity as mentioned in my paper on the fault origin of oceanic deeps (Benioff, 1954). In general, the movement is thus normal to the trend of the coastline. On this basis the curvature of the San Andreas system and the existence of the Garlock Fault are the result of differential expansion of the continent at the Pacific margin—with the northern portion expanding faster.

The movement along the Garlock Fault is sinestral, whereas the movement on the Murray fracture, given by Mason's magnetic surveys, is dextral. Moreover the Mendocino fracture appears to have no expression within the continent east of the San Andreas Fault. I am inclined therefore to the opinion that these oceanic fracture systems are unrelated to the systems shown in my figure [Fig. 32.19, this book]. They are probably older—or at least no longer active since they have no earthquake activity of consequence except in those portions adjacent to the continental margins where it is probably induced by the movements going on there. It would seem to me that if the oceanic fracture systems were closely related to the present radial flux pattern they should be active seismically over most of their lengths.

33.

IGNEOUS AND TECTONIC PROVINCES OF THE WESTERN CORDILLERA

OBJECTIVES

Volcanic rocks cover large parts of the surface of the western United States and, by forming appreciable segments of certain sedimentary sequences, underlie other extensive areas. The Nevadan batholiths are possibly the most voluminous of all rock units. At least three hundred stocks and small batholiths occur in Nevada, Utah, Arizona, Colorado, Montana, and Idaho, and numerous laccoliths, sills, and dikes have been described in the Colorado Plateau, Wyoming, Montana, and Colorado. So much of our attention is focused on the sedimentary rocks that the extensive array of igneous rocks is generally passed by with only incidental reference. It is the object here first to summarize the kinds and distribution of the igneous rocks in the western Cordillera of South and North America, and then second, to find a relation, if any, to the tectonic divisions.

We are always seeking an answer to the deep-seated cause of mountain building, and since the primary magmas are generally thought to have developed in the base of the silicic crust, in the basaltic subcrust, or in the outer mantle shell, it is possible that a careful analysis of the distribution patterns of igneous rocks and their parentage may help us understand the nature of orogeny. This will be the final objective.

CONCEPT OF IGNEOUS PROVINCES

Kennedy's Associations

It has long been recognized that certain regions are characterized by a related assemblage of extrusive and intrusive rocks, and that this assemblage differs from an adjacent one in dominant petrologic types, chemical composition, and nature of extrusion or intrusion. Such a region will here be called an igneous province. The rocks of one province may be relatively uniform in composition such as the basaltic rocks of the Columbia River Plateau, or they may be varied both in mineralogy and chemical composition, such as the olivine basalt–nepheline basalt–melilite basalt–trachyandesite–trachyte–phonolite differentiation series of the San Juan Mountains.

In spite of the striking variation in mineral and chemical composition in these series, it is evident that certain primary magmas are indicated from which the series have evolved either directly by magmatic differentiation or by differentiation along with the assimilation of certain kinds and amounts of country rock. (See Turner and Verhoogen, 1951, for a systematic discussion of the process and problems.)

Professor W. Q. Kennedy of the Scottish Geological Survey postulated in 1933 that the differentiation series and the great basalt fields come from two basic kinds of primary magmas, namely, olivine basalt and tholeiitic basalt. The first is characterized by appreciable olivine and augite, and is commonly alkalic. Kennedy recognized it as the type present in the

oceanic volcanic outpourings and in some of the large basalt fields of the continents. In the second, olivine is generally absent or if present, is subordinate. Pyroxene (hypersthene) is prominent. This is the primary basalt of the majority of plateau or flood basalts, such as in the Columbia River basalt field, generally in the eugeosynclinal assemblages, and to some extent in the andesite complexes of the orogenic belts. The scheme of magmatic descent as he gave it is as follows:

Olivine basalt (Alkalic)	*Tholeiitic basalt* (Calc-alkalic)
↓	↓
Trachyandesite	Andesite
↓	↓
Trachyte	Rhyolite
↓	
Phonolite	

Kennedy also recognized a third magma association which he called the plutonic. This igneous kindred appears to be limited to the cores of orogenic belts, and includes all discordant and concordant batholiths, stocks, and sheet complexes there. It also includes the minor associated aplitic, pegmatitic, and lamprophyric intrusions. The plutonic associations consist almost entirely of granodiorite and granite together with the small amounts of hornblendic, basic, and ultrabasic types. The granodioritic and granitic plutons are generally emplaced after an episode of intense compressional orogeny, but some in places are known to have accompanied the orogeny.

> Many of the rock types possess no effusive equivalents nor has any true subjacent plutonic mass been found within a nonorogenic area. This latter feature alone is sufficient evidence of some fundamental genetical distinction between rocks of the volcanic and plutonic associations.
>
> We know that a granitic liquid can be produced by the fractional crystallization of basaltic magma and, within the volcanic associations, the relative proportion of acid to basic rock types and the chemical composition of the former is consistent with the view that the rhyolites, granophyres and granites of the non-orogenic suites have been formed by high-level differentiation and fractionation of a primary basaltic liquid. This mode of origin applies also to the volcanic associations of the orogenic zones where subordinate quantities of acid lavas are associated with the predominantly basic extrusives.
>
> The acid rocks of the true plutonic associations, however, represent such an enormous bulk of granitic and granodioritic material that it is impossible to conceive of their derivation from a basaltic parent and we are forced to conclude that they must have formed from some primary acid magma . . . (Kennedy, 1933).

Whereas many volcanic associations are believed to have been derived from a basaltic magma which originates by remelting of a universal subcrustal basaltic layer, or by partial melting of the outer mantle, plutonic associations are believed to originate by melting of a downfolded or thickened part of the overlying "granitic" layer. It is commonly stated that such thickening seems possible only where compressional orogeny has caused the base of the silicic crust to extend down into the range of melting.

Turner and Verhoogen's Associations

Following Kennedy, Turner and Verhoogen (1951) define a volcanic association or kindred as one including all igneous rocks, intrusive as well as strictly volcanic, that are genetically related to a cycle of volcanic activity. They emphasize a classification based on oceanic and continental distribution which is as follows:

1. Oceanic associations (for the Pacific)
   a. Olivine basalt–trachyte (Intra-Pacific)
   b. Andesite dacite–rhyolite of marginal island arcs (Circum-Pacific)
2. Volcanic associations of nonorogenic continental regions
   a. Olivine basalt–trachyte–phonolite association
   b. Leucite basalt–potash trachybasalt–trachyte association
   c. Tholeiitic basalts and equivalent quartz diabases
3. Volcanic associations of orogenic zones
   a. Spilite–keratophyre association
   b. Basalt–andesite–dacite–rhyolite association

The Circum-Pacific oceanic association is similar to the continental orogenic basalt–andesite–dacite–rhyolite association. Both are dominantly

andesites and basalts, but a clear relation to either of the parent basalt magma types has generally not been agreed upon or established. Great volumes of andesite are erupted in some orogenic belts with little or no accompanying olivine basalt, and this gives rise to the belief that the roots of the thickened "granitic" crust in orogenic belts may be melted and in part mixed with basalt magma to form andesitic magma directly and even rhyolitic magma at times (Waters, 1955).

Kuno (1954) reports on a volcanic zone on the Izu peninsula southwest of Tokyo, Japan, which is a small part, but perhaps typical, of the Circum-Pacific igneous association. Most of the lavas (basalts and andesites) are characterized by a low $MgO:FeO+Fe_2O_3$ ratio and low alkalies, and also by low normative feldspar rich in An and high normative quartz. However, the lavas of the Omuro-yama group, a small field in the zone, are high in the $MgO:FeO+Fe_2O_3$ ratio and in alkalies. Some of them have a considerable amount of normative olivine, and most of them contain resorbed xenoliths captured from a granitic rock. He concludes that the main rocks of the zone represent various stages of fractionation of a tholeiitic magma, but that the Omuro-yama rocks represent products of contamination by granitic rock. The xenoliths were taken from the wall of the magma reservoir which supplied extrusive lavas, and not from the walls of the conduits, because in order to effect assimilation, the magma must have been in contact with the salic plutonic rock for a considerable time, otherwise only mechanical mixing would have taken place.

The spilites of the orogenic zones are soda-rich olivine-poor basalts, with albite or oligoclase the sole or principal feldspar. Some of the albite in certain flows is secondary. A keratophyre is a sodic trachyte with albite as the principal constituent. Many spilites are pillow lavas and are interbedded with marine sediments; hence probably erupted on the sea floor as submarine flows. The spilites and keratophyres are commonly associated with normal basalts and andesites, and are typical volcanic rocks of the eugeosyncline. Because of this position they are particularly subject to low-grade metamorphism and become the greenschists of the orogenic belts. Waters (1955) regards the spilite–keratophyre association in the Coast Ranges of Washington and Oregon as a tholeiitic province, but Turner and Verhoogen (1951) think the chemical data yet insufficient to establish a clear-cut relation to the tholeiitic or the olivine basalt magma types:

> The spilitic association, whatever its relation to the basaltic kindreds, is one of striking individuality maintained in widely scattered provinces of all ages and recognized wherever the rocks of geosynclinaal terranes have been petrographically investigated (Turner and Verhoogen, 1951, p. 205).

The olivine basalt–trachyte–phonolite association is displayed in places in the Rocky Mountains, particularly in moderately deformed belts of Laramide orogeny. It is an extensive differentiation series ranging from olivine basalt to basanites to trachybasalts and trachyandesites to phonolites. The members generally have alkaline affinities. Within a single volcanic episode hundreds of flows together with much pyroclastic material may be erupted from numerous centers to form a continuous field 50 to 75 miles across. Intrusive sills, laccoliths, plugs, and dikes are a minor part of the field. Xenoliths are commonly conspicuous in the flows and several authors believe the original olivine basalt magma was contaminated by reaction solution (fusion) of the wall rock. The type of wall rock and the amount assimilated determines to a large extent the course of differentiation of the magma. This general association is represented by the San Juan volcanic field (Larsen and Cross, 1956) and probably other fields in Colorado and New Mexico.

The leucite basalt–trachybasalt and trachyte association in the western United States is represented by the feldspathoid, alkali-rich rocks of the Colorado Plateau, Leucite Hills and Black Hills in Wyoming, and the well-studied region of central Montana (Larsen, 1940).

The association called tholeiitic basalts and equivalent quartz diabases are the flood basalts of such volcanic fields as the Columbia River Plateau. The most striking characters are the enormous volume, wide extent, and uniform composition of the basalt sheets.

Tyrrell's Tectono-Igneous Cycle

Emphasizing the tectonic and time aspect of petrographic provinces Tyrrell (1955) has proposed the following tectono-igneous cycle." It applies to the complicated region of northwestern Europe consisting of

"three ancient orogens welded onto the Scandinavian-Baltic shield," and particularly to the Scottish Paleozoic.

Diastrophism	Kindreds	Locus
I. Geosynclinal phase	1. Ophiolitic kindred	In orogen
II. Orogenic phase (with two or three subphases)	2. Granodiorite–andesite kindred	In orogen
III. Post-orogenic phase (with two subphases)	3. Trachybasaltic kindred 4. Quartz dolerite kindred	In kratogen

Proposed Classification of Provinces

In the western United States, certain igneous rock associations stand clearly apart from others. Discussions generally center about such striking igneous provinces as the Cascade Mountains, the San Juan Mountains, or central Montana, yet no one has published a map of the entire western United States on which are grouped the many volcanic fields and plutons into igneous provinces. Several emphasize the transitional and elusive nature of boundaries, and this is certainly realized when one attempts to draw them. The main goal of this chapter is thwarted, however, if the petrographic provinces are not mapped and compared with the tectonic provinces.

In struggling with the problem, difficulties in two categories arise. First, in the provinces of extensive basalt outpourings a distinction between rocks of the olivine basalt kindred and the tholeiitic kindred is commonly obscure. The problem is met with specifically in classifying the Malheur and Snake River basalt fields. Second, in the alkalic and calc-alkalic "provinces" of the Rocky Mountain states, the boundaries of the numerous subdivisions suggested in the literature are generally impossible to fix or map. Second, the main kind or kinds of rock present is generally a characteristic feature which can be mapped objectively, whereas the kindred represented may be controversial.

A classification believed better suited for tectonic studies is as follows. It will serve as a guide in the discussion of the igneous rock provinces of the western Cordillera of the Americas, and is especially adapted to the western United States.

A. Basalt provinces
   1. Oceanic (mostly olivine basalts)
   2. Continental flood and cinder cone fields (both olivine and tholeiitic basalts)
B. Andesite provinces
   1. Eugeosynclinal (mostly tholeiitic basalts and andesites-spilites and keratophyres characteristic)
   2. Volcanic arcs
   3. Orogenic belt (post-batholithic volcanic fields)
   4. Stratovolcanos of continental margin
C. Trachyte and phonolite provinces
   1. Alkalic (leucite basalt–trachyte–phonolite group)
   2. Calc-alkalic (olivine basalt–phonolite association, also andesite and rhyolite)
D. Latite-monzonite provinces
E. Basalt-rhyolite provinces
F. Grandiorite-granite batholithic provinces
   1. First cycle
   2. Second cycle

The petrologic terms basalt, andesite, latite, trachyte, and phonolite are used to denote the main type of rock of the province. In the andesite provinces especially, differentiation products are common as well as olivine and tholeiitic basalts. The basalt-rhyolite provinces specify those in which the intermediate to subacid differentiates are dominant.

Evident Tectonic and Igneous Cycle

In the orogenic belts that form the margins of the continents, such as exemplified by the Sierra Nevada of California and the Acadian belt of New England, the main events follow a fairly consistent pattern or cycle. The one given below is modeled after Turner and Verhoogen, (1951), but with additions and modifications as seen necessary from a study of the Cordillera of South and North America.

1. Eruption of dominantly basic (spilitic, keratophyric, basaltic, and andesitic) lavas during the eugeosynclinal phase.

2. Injection of ultrabasic and basic plutonic intrusions into the eugeosynclinal sediments and volcanic rocks which are almost constantly being disturbed by episodes of folding.

3. The climactic folding and dynamic metamorphism of the eugeosynclinal rocks. In the South American Andean system the folding seems to have been mostly late Paleozoic, preceding the Late Cretaceous batholiths by a long time. Much eugeosynclinal volcanic rock accumulated between the metamorphism and the batholithic intrusions. In the Sierra Nevada a series of orogenic phases stretching at least from the Devonian to the Cretaceous preceded the batholithic intrusions. In Late Jurassic time intense folding and low-grade regional metamorphism climaxed the train of disturbances. Batholithic intrusions followed immediately in Late Jurassic and again, most voluminously, in Mid-Cretaceous. In the Acadian belt of New Hampshire, specifically the White Mountains, early folding and thrusting resulted in regional low-grade metamorphism, then followed the main batholithic intrusions with accompanying medium and high-grade metamorphism, and finally a second episode of thrusting. The three stages occurred within late Devonian time.

4. Emplacement of the great granodioritic and granitic batholiths into the folded and metamorphosed complex. Some batholiths are involved in the folding, but the great bulk of the plutonic rock is post-folding in age. Each great batholith is commonly composed of a number of individual plutons with each having a slightly different composition. They range from diorite to granite with granodiorite the most abundant. In the Sierra Nevada the sequence of intrusions seems to have occurred over a period of 18 m.y.

5. An extensive episode of erosion in which the batholithic rocks are exposed, with the development commonly of adjacent longitudinal basins of sedimentation.

6. Renewed volcanism with the building of great lava and pyroclastic fields, chiefly andesitic, on the folded and metamorphosed batholithic belt. These are the Tertiary volcanic fields of the Andes and of the Sierra Madre Occidental of Mexico, and probably the Mississippian (?) Moat volcanics of the White Mountains. In places latites may be very abundant.

7. Following shortly the post-batholithic eruptions and probably part of the same renewed igneous activity are new batholithic intrusions which in places reach up to the volcanic accumulations and intrude them. Examples are the White Mountain magma series of the White Mountains, the post-volcanic batholiths of the Cascade Range of Washington, and the imposing belt of mid-Tertiary batholiths of western Sonora. This is the second cycle batholithic province of the proposed classification above.

8. A late volcanic activity occurs in segments of the orogenic belt and results in the building of a majestic row or belt of stratovolcanoes, or an extensive field of basaltic flows and cinder cones.

In the western United States the broad Paleozoic miogeosyncline and shelf, and the superposed Mesozoic basins and Laramide belts of deformation, are replete with igneous rocks of the trachyte, phonolite, and latite associations. When compared with the Andean, Mexican, and Canadian Cordillera, the wide and complex western United States Cordillera is an exception. Tectonic provinces like the Great Basin, Colorado Plateau, and the Wyoming and Colorado Rocky Mountains belt of orogeny are either not developed to the north and south of the United States or are reflected in a narrower or restricted way. It is the object of the following pages to review the petrographic and igneous provinces of the western Cordillera of South and North America by using the above depicted concepts in an attempt better to understand the process of orogeny.

34.

IGNEOUS AND TECTONIC PROVINCES IN SOUTH AMERICA

CHILE AND ARGENTINA

Geosyncline

Two references are most significant for a general understanding of the Andean geology of South America, viz., *Handbook of South American Geology, Geol. Soc. Am. Memoir 65,* 1956, edited by W. F. Jenks, and *Bau der Sudamerikanischen* Kordillera (Gebrüden Borntraeger, Berlin) by Heinrich Gerth, 1955. The Cordillera of Chile and western Argentina marks essentially the site of a previous geosyncline, particularly the eugeosynclinal division. (See Fig. 34.1.) Its Paleozoic history is not well known, but as far as a great igneous and geosynclinal cycle is concerned we may start with the late Permian and early Triassic, when continental conditions probably prevailed. During this time voluminous extrusions of keratophyre and quartz porphyry occurred.

> . . . These extrusions are pierced by granites which are the intrusive phases of the lavas. Later, the sea advanced from the west, eroding the volcanic rocks and depositing a transgressive series which has at its base the products of the destruction of the volcanics which in turn pass upward into shale with a marine fauna. This transgression marks the beginning of the Andean geosyncline. Later the keratophyre extrusions were renewed, with a more basic composition than previously, and flows partially filled the marine basin. Plant-bearing shales were deposited. But all these episodes were transitory because the ocean transgressed again during the Norian (late Triassic), with the deposition of thick layers of shale. Later, continental conditions returned, perhaps because of tectonic movements whose nature is unknown. The topography formed was then destroyed during the Rhaetian (latest Triassic) when a surface was prepared for the Liassic (early Jurassic) transgression (Cristi, 1956, p. 197).

The Triassic volcanic rocks are over 12,000 feet thick in the Frontal Cordillera of Mendoza but thin toward the east.

Late Triassic volcanism continued into early Jurassic time but the distribution of the eruptives is possibly limited to southern Atacama and northern Coquimbo.

> . . . In the rest of Chile andesitic volcanics seem to be lacking in the Lias. However, in the Coast Range of Aconcagua and Valparaiso, the Upper Lias sediments contain thick keratophyre flows and tuffs. Apparently similar conditions are found in the Argentinean Cordillera. It is interesting to note that this type of extrusion is not known in the region north of Atacama; this proves that the keratophyre extrusions, which probably began during the Lower Triassic in an area of enormous size, become more and more restricted. At the same time acidity of the flows diminished. This phase of volcanism ended in the late Liassic.
>
> During the Upper Dogger (mid-Jurassic), andesitic extrusions covered almost all the area occupied by the western part of the geosyncline; but, at least in the Coast Range and the central zone, it seems that before these lavas were deposited many important tectonic movements occurred, possibly in the form of block faulting, since in some places the deposits lie on Triassic and in others on Liassic formations.
>
> We know little about the mechanism of these extrusions, but judging by some masses of andesite which pierce the Triassic or Liassic, and by the abundance of pyroclastic materials in the series, probably they were produced by volcanoes of the central type, which must have been elevated above the sea

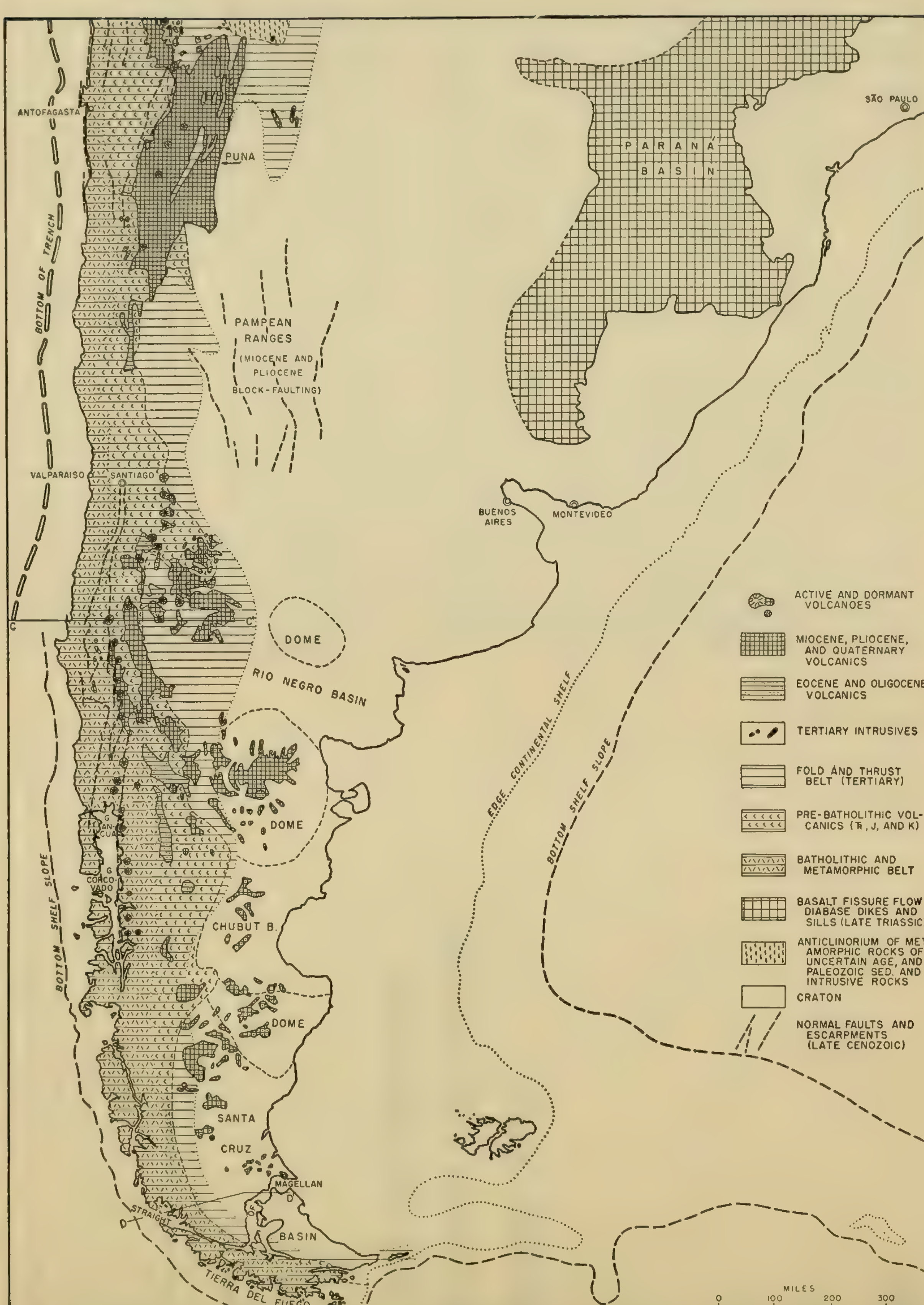


Fig. 34.1. Major tectonic and igneous units of the southern half of South America.

bottom, spreading their lavas partly as submarine and partly as terrestrial flows. These lavas and pyroclastics repeatedly filled large areas of the basin. They were subjected to marine erosion, and conglomerate and limestone were deposited on the marine terraces. Finally the filling became so thick that much of the basin acquired continental characteristics, except in the eastern zone, where sediments continued to be deposited until early Malm (Upper Jurassic) time (Cristi, 1956).

The earlier keratophyres gave way to Jurassic andesites which occurred as flows, "porphyritic" tuffs and welded tuffs, and andesite breccias and conglomerates (boulders are generally andesite porphyry). According to C. Lomnitz of the University of Chile (personal communication) some of the keratophyres mentioned in the literature are probably spilites, and pyroclastics predominate over lavas.

Volcanism continued into earliest Cretaceous time with the accumulation of andesite breccia conglomerates and red sandstones derived from the volcanic rocks. Marine sediments are extensive along the Chilean-Argentinian border, and it is thought that the intertonguing volcanic rocks graded into a volcanic chain along the western coast.

The Santa Cruz basin (also called Magellan geosyncline) on the south lacks Late Triassic volcanic rocks, but in the Jurassic intense volcanism broke out there, and a thick series of keratophyre and andesite flows accumulated. These are called the "Serie Porfirica." (See Fig. 34.1 and section D–D′, Fig. 34.5.)

. . . The Serie Profirica of the Magellan region has frequently been likened to the Triassic keratophyres of Central Chile and Argentina. Although the two series show great petrographic similarity, they are not synchronous since according to modern studies by Argentinian geologists, the Serie Porfirica starts during the Jurassic and ends during the Lower Cretaceous. Another important difference between the two extrusive aggregates is that the Mesozoic extrusions of the Andean geosyncline changed from quartz-keratophyres to andesites, whereas in the northern part of the Magellan geosyncline the acidic character was maintained during the whole interval, and only in one place do a few unimportant andesites appear.

On the other hand, formation of both basins was preceded by extensive keratophyre eruptives.

The rocks above the Serie Porfirica in the Magellan region consist of dark fine-grained sediments with a phyllitic aspect and include marly clay shales and graywackes. Radiolaria are abundant in the lower beds. A few basic dikes cut the lower strata. In the Cordillera of the Brunswick Peninsula and Tierra del Fuego this series must be several thousand meters thick (Cristi, 1956).

These seems to be little development of a miogeosynclinal division of the geosyncline; the volcanic sequences and interbedded sediments thin to the shelf area of the foreland with volcanic rocks making up part of the much thinner sequences.

Batholithic and Metamorphic Belt

As seen on the map of Fig. 34.1 the entire coastal zone of Chile is made up of a belt of metamorphic and batholithic rock. From Valparaiso southward a metamorphic rock zone stretches along the shore almost to the Straits of Magellan. The batholithic zone in this segment lies inland or east of the metamorphic belt but gradually transgresses the metamorphic zone and comes to the coast just north of the Straits of Magellan. As the combined belt veers eastward through Tierra del Fuego, the metamorphic belt appears on the inside or to the northeast. The reality of a great batholith is not questioned. In some cross sections it is shown underlying the whole Cordillera, from the Coast Range through the high Andean system. On the *Geologic Map of South America* (1950) it is 75 miles wide just east of the Gulf of Corcovado and is almost continuous from Valparaiso southward through Tierra del Fuego. It has thinned to about 25 miles at Valparaiso and from there extends as a narrow belt another 400 miles northward. It then becomes discontinuous and is represented by scattered intrusions into southern Peru. From Valparaiso northward it is intrusive into the eugeosynclinal sediments, mostly Late Triassic and Jurassic volcanics.

The relation of the batholiths to the metamorphic rocks has not been clearly established. In places the metamorphic rocks are intensely folded Paleozoic strata, but the orogeny may be Permian or earlier, at least in places, and not an immediate prelude to the later batholithic intrusions. Extensive gneissic zones in the Coast Range near Valparaiso are probably migmatites of the batholith.

Brüggen (1934) has very carefully analyzed the phenomena at the contact between the batholith and the country rock, which is generally the Serie Porfirica, and he has concluded that the gneissic appearance of the

batholith in places near the contact is due to migmatization. No doubt, in numerous places a certain amount of migmatization has taken place, but generally the gneissic aspect consists of a folded primary structure, injected by veins during the latest stages of magmatic consolidation.

Both the gneissic appearance and the migmatization are much stronger in the Coast Range, probably because in this range the lower levels, where magmatic phenomena could develop with greater efficiency, are more accessible to observation. For this reason the batholith has until recently been considered very old. But Brüggen demonstrated through his analysis of numerous outcrops that the intrusion cannot be older than Early Cretaceous, because the strata of this age which are near the batholith are always affected by thermal metamorphism. Besides, the Paleozoic and Triassic conglomerates, which are relatively abundant in Coquimbo, contain no pebbles that could have been derived from the Andean batholith; the pebbles are aplitic granite containing albite or microperthite. Such rocks are subordinate in the batholith, which is mostly tonalite and granodiorite. However, in the Coast Range the first phases of the intrusion might well have been tied to the orogenic movements which occurred in the Jurassic (Cristi, 1956).

Although the great piles of volcanic rock were intruded, a zone of the eugeosyncline along the east side of the batholithic zone was left free of intrusions. The layers of volcanic rock, here many thousands of feet thick, were tilted so as to dip eastward and from this great monocline in the region east of Santiago the high peaks of the main Andean Cordillera were carved.

Petrographically the main batholith ranges from granodiorite to tonalite. Fairly extensive bodies of granite are known, especially aplitic granite. Gabbro and hornblendite are listed as present and are said to be basic derivatives, presumably of the parental dioritic magma. Diorite porphyry and dikes of lamprophyre, aplite, and especially kersantite and spesartite are mentioned. True pegmatites are uncommon (Cristi, 1956).

PERU, BOLIVIA, ECUADOR, AND COLUMBIA

Geosyncline

Geosynclinal sedimentation is known to have become established in Middle Ordovician time in Peru approximately in a north-south basin in the site of the present Andes. The basin extended, at least, as far west as the present shoreline. Silurian strata have not been observed but Middle Devonian strata are known in places. (See map, Fig. 34.5 section B–B′.)

In central Peru, Middle Pennsylvanian beds rest disconformably on the Middle Devonian. In southern Peru, Permian beds rest disconformably on older Paleozoic hard, micaceous shale. Again in central Peru, Permian conglomerate and red beds rest in strong angular unconformity on older Paleozoic rocks in various conditions of metamorphism. In southern Peru, marine Permian (and possibly Carboniferous) beds cover contorted and metamorphosed older formations. Mississippian of Pennsylvanian continental deposits rest unconformably on older rocks in northwestern and central Peru. Absence of known Upper Devonian and marine Mississipian strata in Peru is a further suggestion of orogeny and uplift beginning at the end of Middle Devonian time.

In central Peru late Paleozoic orogeny must have begun at the close of the Middle Pennsylvanian. Continental clastics and volcanics of Permian age here rest disconformably on the Pennsylvanian sequence (Jenks, 1956).

Permian time was marked in places by marine transgressions and by the deposition of 2300 feet of "red beds and conglomerates" apparently of continental origin. In central and southern Peru the Permian volcanic rocks attain great thicknesses.

The close of the Paleozoic in Peru was marked by strong orogeny. Permian granites and associated quartz veins cut thick folded and faulted Paleozoic metamorphics in northwestern and probably in southern coastal Peru. There was intense volcanic activity and general emergence of the Andean zone in upper Permian time. Evidently the whole of Andean Peru was land from some time in the Upper Permian until at least the beginning of the Upper Triassic (Jenks, 1956).

The history thus related is responsible for the belt of metamorphic rocks shown on the map, Fig. 34.2, which stretches from northernmost Chile to northwestern Peru. Its present borders are probably due to later orogeny. A similar belt of metamorphic rocks forms the Cordillera Oriental of Ecuador. Its paragneisses and paraschists probably represent Paleozoic and perhaps some Precambrian sediments deposited in an Andean geosyncline.

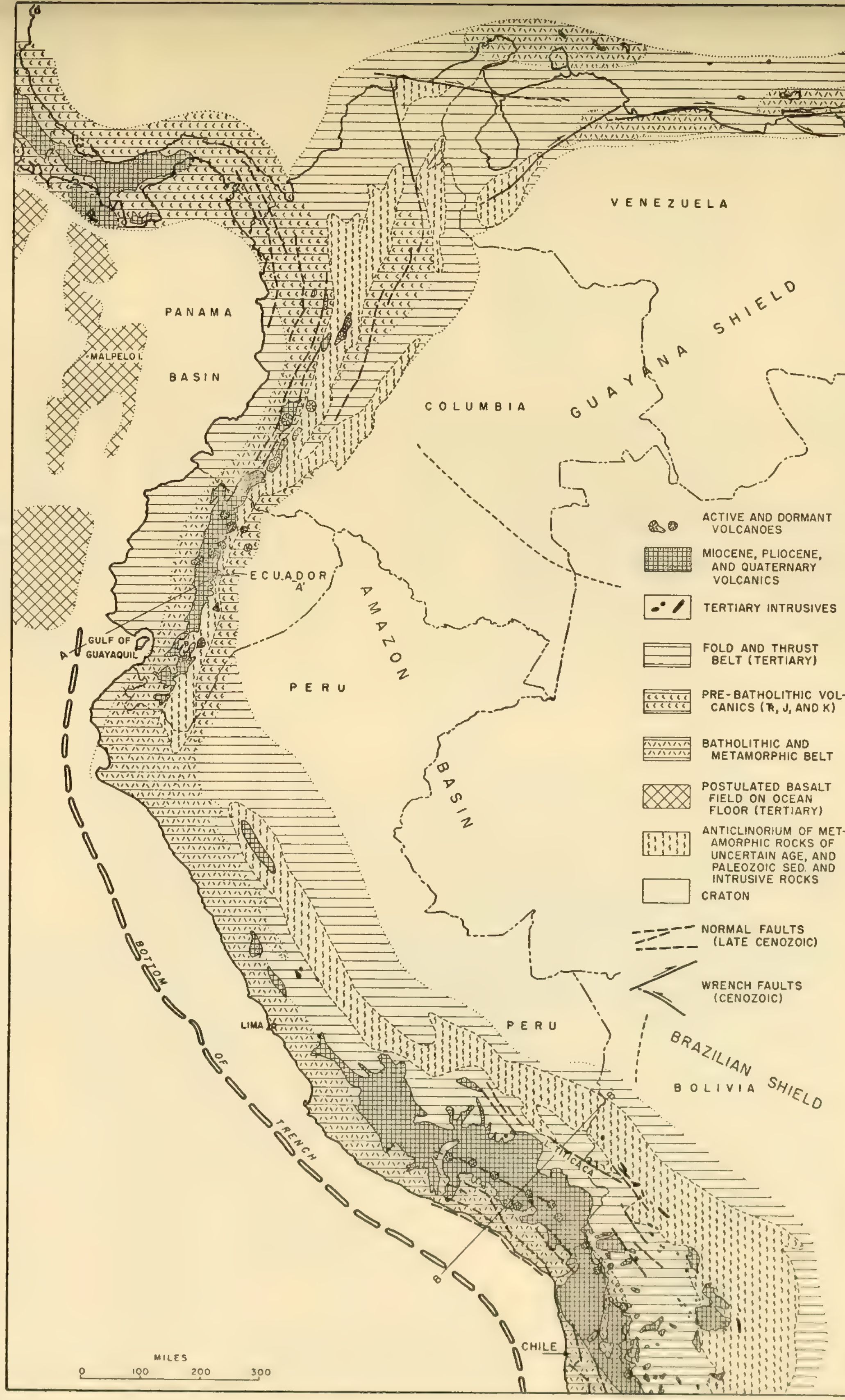


Fig. 34.2. Major tectonic and igneous units of the northern half of the Andean Cordillera. Details of the Venezuelan belt not included.

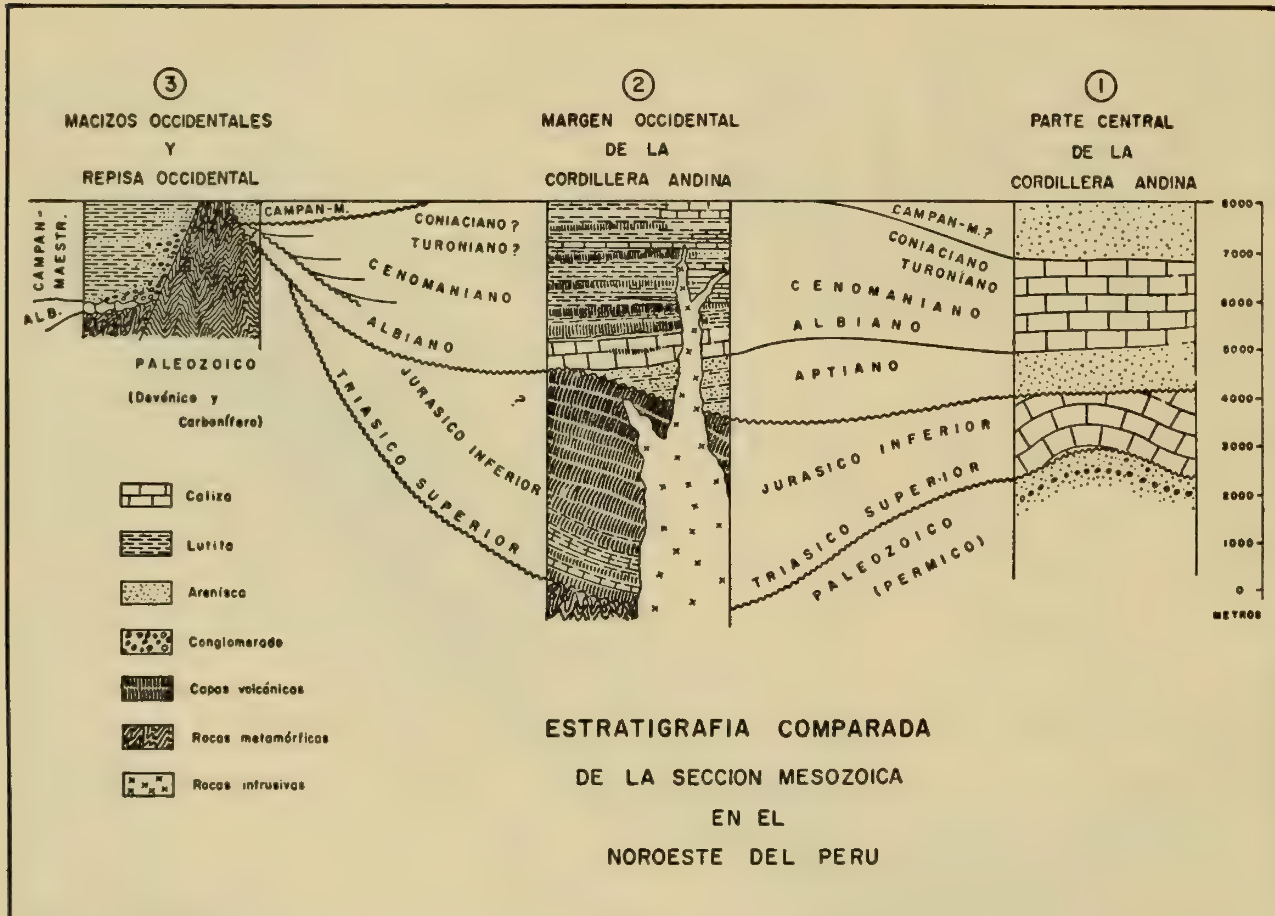


Fig. 34.3. Comparative sections of the Mesozoic of northwestern Peru. Reproduced from Fischer, 1956.

Middle Paleozoic to early Mesozoic disturbances may well have contributed to the metamorphism of the present day crystalline rocks, but there are more hiatuses than basic data . . . Some of the orthogneisses probably represent pre-Jurassic intrusions (Jenks, 1956).

The Mesozoic history of the Andean geosyncline in Peru and Ecuador is in general like that of Chile, and its main stratigraphic elements are described by Fischer (1956) whose illustrations are reproduced as Figs. 34.3 and 34.4. The pre-batholithic Mesozoic volcanic rocks extend northward from Chile into southern Peru but there wedge out against the coast. They reappear in northwestern Peru and Ecuador. The map of Fischer (Fig. 34.4) and the map of the Andes of Peru, Ecuador, Columbia, and Bolivia (Fig. 34.2), which was taken from the *Geologic Map of South America* (1950), represent the distribution of the Mesozoic pre-batholithic volcanic rocks (Jenks, 1956). The concept of a volcanic archipelago and eugeosyncline on the ocean side (west), and then the miogeosyncline on

the mainland side is portrayed. The volcanic arc and eugeosyncline are about 200 miles wide. In California and western Nevada the same has a width of about 400 miles. Permian volcanic rocks are abundant in both places.

Batholithic Belt

The batholithic belt extends up the coast from Chile to northwest Peru, and the great plutons form much of the western slopes and foothills of the Cordillera Occidental. In northern Peru the batholithic belt swings northeast across Peru, following the Mesozoic geosyncline. According to the *Geologic Map of South America* (1950) the plutons are separate and small in Ecuador, and in Colombia the belt becomes so inconspicuous that its identity, at least on the map, is problematical.

The rocks of the great coastal batholith of Peru are, in order of abundance, granodiorite, tonalite, granite, and diorite. Others in small volume are quartz monzonite, monzonite, syenite, and gabbro. Wide, deeply eroded parts of the batholith appear to be fairly homogeneous in composition, but apically truncated parts show a wide range of petrologic types (Jenks, 1956).

The main part of the batholith was intruded, apparently, in early Upper Cretaceous time. Lower Senonian and even Turonian strata have been intruded and metamorphosed, but younger ones have not been affected. Lower and Middle Cretaceous rocks in Peru contain abundant volcanics, but when the batholithic intrusions ocurred, no further volcanism is recognized through the Maestrichtian, Danian, and Paleocene to the late Eocene. Abundant volcanic rocks appear to be largely Miocene, Pliocene, and Quaternary in age.

Anticlinorium of Pre-Mesozoic Rocks

A belt of Paleozoic and Precambrian rocks extends from the Argentine border northwestward through Bolivia and lengthwise through Peru almost to Ecuador. It is lacking for about 100 miles and then at the southern Ecuadorian border commences again and extends through Ecuador and nearly through Colombia. In disconnected areas it is present in western Venezuela. (See map, Fig. 34.2, and cross section, Fig. 34.5.) In

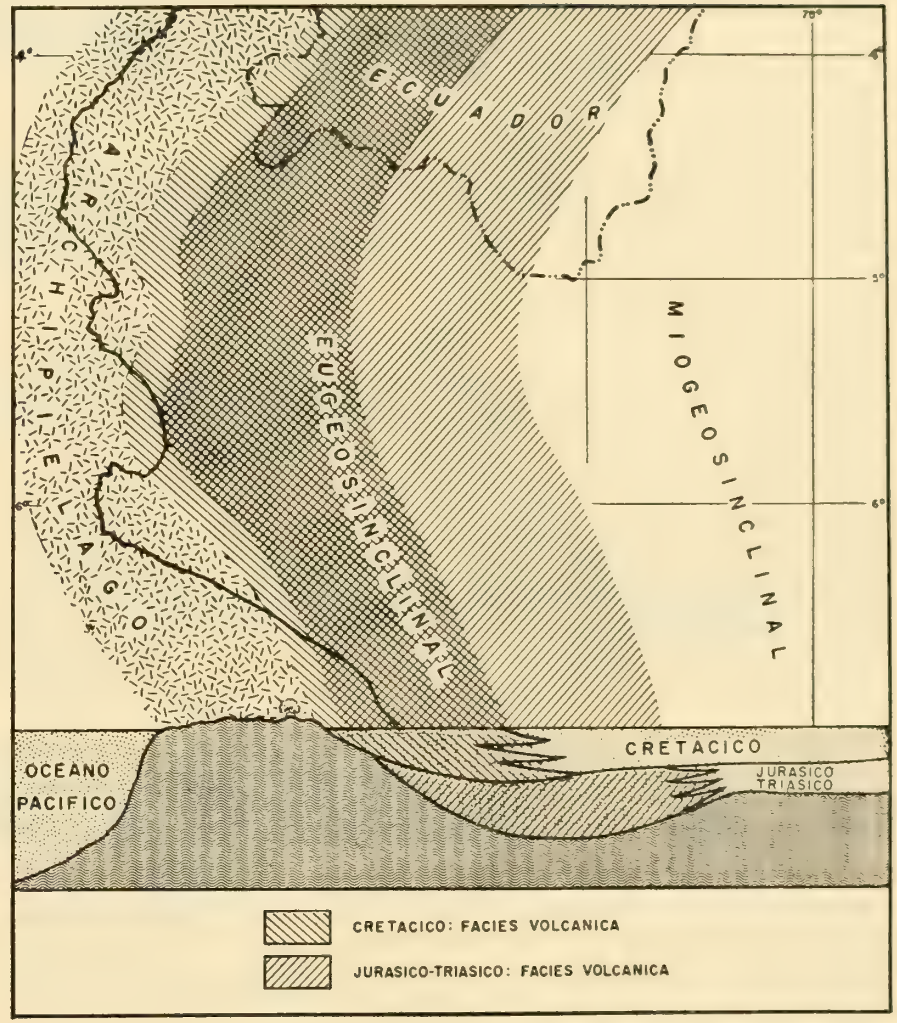


Fig. 34.4. Idealized restoration of Mesozoic sedimentary and tectonic divisions in northwestern Peru. Reproduced from Fischer, 1956.

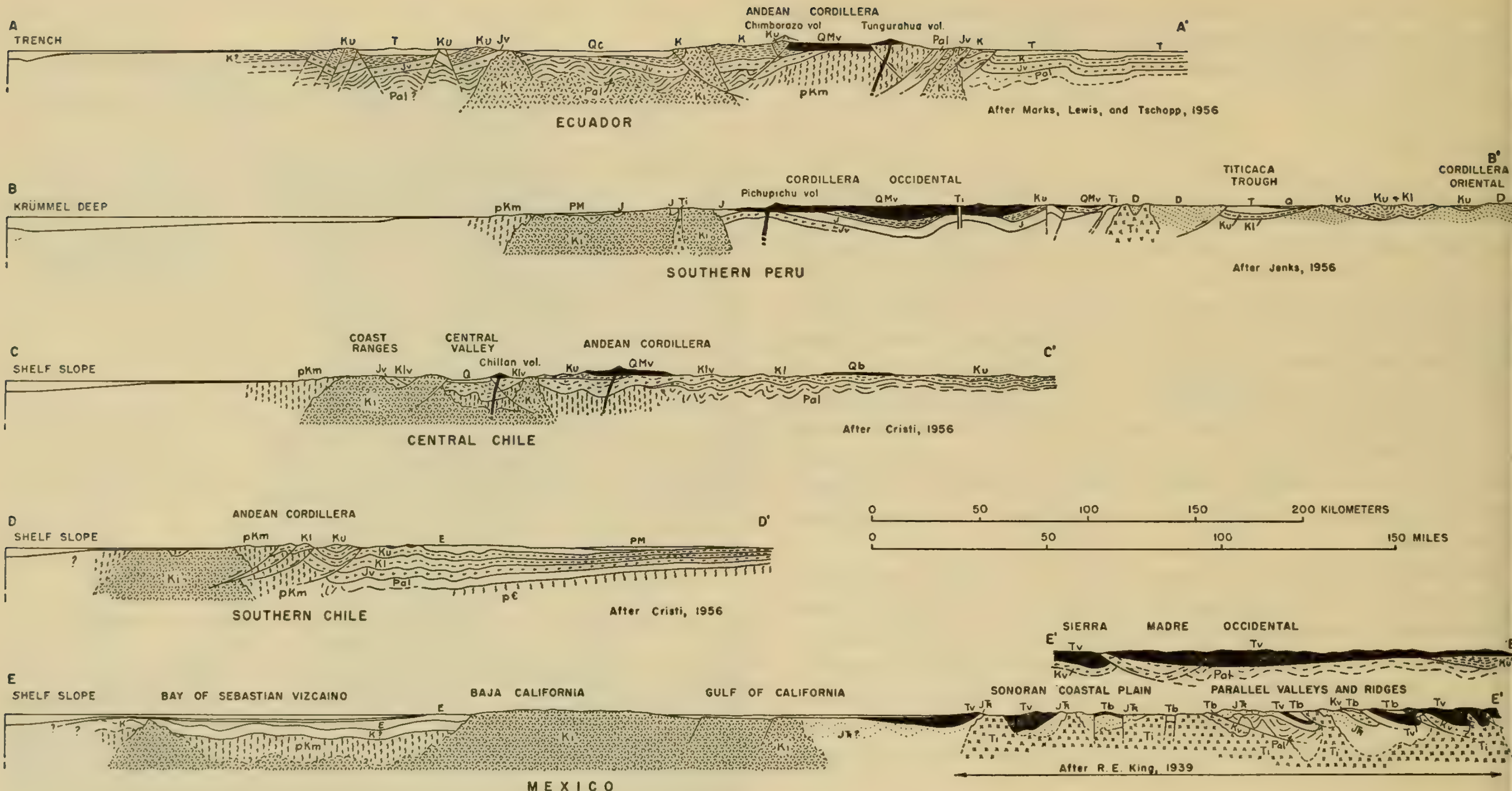


Fig. 34.5. Cross sections of the western Cordillera of South America and Mexico showing major igneous and sedimentary units. See maps, Figs. 34.1, 34.2, and 35.1 The vertical scale is about twice the horizontal scale but, even so, some of the thicknesses of the stratified units are undoubtedly too great. Interpretations at depth have been made which are not shown on original sections of authors cited.

PЄ, Precambrian; Pal, Paleozoic; D, Devonian; JŦ, Jurassic and Triassic strata; J. Jurassic strata; Jv, Jurassic volcanics; pKm, pre-Cretaceous strata; Klv, Lower Cretaceous volcanics; Kl, Lower Cretaceous strata; Ku, Upper Cretaceous strata; K, Cretaceous strata; Ki, Cretaceous batholithic intrusives; Kv, Cretaceous volcanics; T, Tertiary strata; Tb, Tertiary basin beds; Tv, Tertiary volcanics; E, Eocene strata; QMv, Quaternary to Miocene volcanics; Ti, Tertiary intrusives.

Bolivia the broad Paleozoic area comprises the eastern and central Cordilleras, and the strata are folded but evidently not enough to produce much metamorphism. In Peru the belt is narrower with considerable folding, faulting, and metamorphism, and may include Precambrian rocks. It also includes several large intrusions, one of which is mentioned as a granite (Jenks, 1956).

In Ecuador the crystalline rocks are highly metamorphosed and form the backbone of the Cordillera Oriental (also called Eastern Andes and Cordillera Real). The types are orthogneiss and paragneiss, mica, garnet schists, also amphibolitic, sericitic, talcose, and graphitic schists, phyllites, and some quartzites and calcareous slates. Also prominent are metamorphosed granodiorites. Minor amounts of metamorphosed syenite and low-quartz granite are noted. The belt of crystalline rocks is flanked on the east by little or non-metamorphosed Paleozoic and Mesozoic strata, with associated volcanic rocks and Mesozoic (?) granites. The complex is thrust eastward at its eastern margin. (See section A–A′, Fig. 34.5.)

Between the Cordillera Occidental, or batholithic belt, and the Cordillera Oriental, or older metamorphic belt, is the intercordilleran depression. It may be compared to a huge graben bounded on the east and west by fault zones which dip away from the graben at fairly high angles. Beginning in Miocene time, as far as known, the graben has had large amounts of volcanic materials poured into it. Faulting is believed to have continued intermittently during the accumulation of the volcanic rocks.

> As the two cordilleras have risen relative to the intercordilleran depression, the volcanoes in and bordering the depression have filled it with vast quantities of predominantly andesitic pyroclastic and flow rocks. At the same time, heavy rainfall and melt water from the snow and ice-clad heights of the cordilleras have eroded the depression-facing slopes and deposited the resulting clastic sediments in the depression. The huge area flooded by this volcanic and—to a lesser degree—clastic fill makes up by far the greater part of the intercordilleran depression (Lewis *et al.*, 1956).

The prominent anticlinorium of Paleozoic and Precambrian rock is for most of its length bounded by reverse faults, and is interpreted to be thrust over the flanking basin sediments on the east and also over the rocks on the west such as the intercordilleran depression in Ecuador and the Lake Titicaca trough at the Peru-Bolivia border. These faults delimit the raised zone of older metamorphic rocks, but do not mark the original width of it.

In Figs. 34.3 and 34.4 highly folded Paleozoic strata are shown to form the core of the postulated Mesozoic volcanic archipelago on the Pacific border of the continent. This picture is built from a few small outcrops, but nonetheless it is as logical a foundation for the Mesozoic volcanic effusives as any. It must be concluded that the width of the belt of Paleozoic folding and metamorphism is much wider than that exposed in the anticlinorium.

POST-BATHOLITHIC VOLCANIC ROCKS

Age Relation to Batholithic Belt

Following the batholithic intrusions and the accompanying folding and faulting a long cycle of erosion removed much of the roof rock and in places cut deeply into the plutons. The surface developed over much of the adjacent Cordillera also. Upon this extensive erosion surface new effusives were spread. The plutonic cycle occurred in most places during Mid- and Late Cretaceous times, and the earliest eruptives are late Eocene, but the main volcanic activity in most places did not start until the Miocene. There was a lapse of time, then, of about 40 m.y. between the plutonic cycle and the beginning of the volcanic cycle.

Areas of Volcanic Rocks

The volcanic accumulations may be grouped in three divisions: (1) between Santiago and the Straights of Magellan; (2) southern Peru, Bolivia, northern Chile, and northwestern Argentina; (3) Ecuador and southwestern Colombia. All are confined to the general Cordillera except in Southern Argentina where flows occur on the foreland. Each of the three areas support a magnificent belt of active or dormant volcanoes in addition to extensive volcanic fields. (See maps, Figs. 34.1 and 34.2.)

The central division is the largest in areal extent and undoubtedly the largest in volume. It occurs around the bend of the great Cordillera from

the northerly trends of Chile and Argentina to the northwesterly trends of Bolivia and Peru. The southern division is notably an assemblage of individual smaller fields.

Although the eugeosynclinal, batholithic, and fold belts are continuous from Tierra del Fuego to Colombia, the Cenozoic volcanic fields are not.

Spatial Relations to Older Belts

In a broad way the volcanic fields lie between the Cordillera Oriental and the Cordillera Occidental, and in part fill a graben whose sinking between the two linear relief elements was about contemporaneous with the eruptions. The volcanic rocks, however, spread over both adjacent Cordillera extensively in places, and in northernmost Chile extend westward across the batholithic belt to the coast. In the Puna de Atacama region of northern Chile opposite Antofagasta, the extrusions are entirely east of the batholithic belt and mostly on the eugeosynclinal strata free of batholithic intrusions. They spread eastward, also, to the deformed miogeosynclinal and shelf sediments. Possibly 800 active and inactive cones exist in this large field and seem to be arranged in several rows. A few scattered fields are on the Precambrian and Paleozoic anticlinorium to the north.

In Ecuador and southwestern Colombia the main volcanic field fits rather snugly in a graben between the batholithic belt on the west and the anticlinorium of older metamorphosed rock on the east. Some of the great stratovolcanoes have vents through the cordilleran rocks on either side, however, and have built considerable volumes of ejecta on these foundations beyond the faults that bound the graben.

The southern division of volcanic fields is generally east of the batholithic belt on the batholithic-free Mesozoic volcanic rocks, and as mentioned, a number of isolated fields lie on the miogeosyncline and shelf areas of the foreland. This division is singular in that the row of great stratovolcanoes is mostly in the batholithic belt and not a part of the volcanic fields. At the north end of the division, the row cuts acutely into the eugeosyncline, and several vents are offset sufficiently far east so that they are in the miogeosyncline (see map, Fig. 34.1). South of Santiago numerous volcanoes have been active in recent years. The great isolated volcanoes or groups of volcanoes are spaced at about 30- to 40-kilometer intervals in this part of the zone.

Composition

The flows associated with active volcanoes in Chile are mostly basalts, ranging from hypersthene basalt in the oldest flows to olivine basalt in the more recent, with the exception of Calbuco, which still erupts hypersthene basalt (Cristi, 1956, p. 213).

The volcanic fields of the foreland in Argentina are nearly all basalt flows of Pliocene-Quaternary or Quaternary age. The Eocene and Oligocene volcanic rocks in the cordilleran region are andesites and dacites, and overlying Miocene flows are basalts.

The great Puna field consists of augite and hypersthene andesite with the latest flows of olivine basalt. Rhyolite is also reported.

The Tertiary and Quaternary volcanic rocks of southern Peru range in composition from basalt to rhyolite, with andesite, trachyandesite, and trachyte very abundant. Cutting the extrusives are numerous small stocks of diorite, monzonite, quartz monzonite, syenite, and dacite porphyry.

The volcanic rocks of Ecuador and southeastern Colombia are dominantly andesitic pyroclastic and flow rocks.

It may be concluded, therefore, that andesites are the most abundant of the Cenozoic volcanic rocks which appear within the Cordillera, with olivine and hypersthene basalts probably next in abundance and also usually latest in the eruptive sequence. In southern Peru the trachyte volcanics and the monzonite and syenite stocks are unusual because of their high alkalic content.

Relation to Graben Faulting

Both Gerth (1955) and Cristi (1956) emphasize the relation of faulting to volcanism, or more generally stated, to "recent tectonic depressions." In the entire Andes only where a well-developed longitudinal valley exists do volcanoes occur. This concept relates specifically to the rows of active and dormant stratovolcanoes. In the Ecuador division, however, the entire field is fairly closely tied to the graben faulting which here has been interpreted as of compressional nature.

The broad and somewhat irregular field of Southern Peru, Bolivia, and northern Chile and Argentina is less positively tied to faults even though the modern volcanoes seem to be. The Puna field could perhaps be developed over a Basin and Range type of faulted terrane, judging from the several rows of volcanic vents there.

Extending from Santiago southward for nearly 1000 miles is a depression that separates the Coast Ranges from the Andean Cordillera. This is called El Valle Central, and is believed to be a complexly faulted graben. The zone of active and dormant stratovolcanoes is almost exactly commensurate in length with the depression, and in the central and southern part the volcanoes follow closely the eastern side or are within the graben. At the north end they occur about 60 miles to the east of the graben.

The fault zones do not bear the same relation everywhere to older tectonic units. In Ecuador the graben occurs between the batholithic belt and the older anticlinorium. In southern Peru the fault zone is mostly within the batholithic belt or along its east side, and in the Puna de Atacama it is developed on the nonintruded eugeosynclinal strata.

The great, tilted, fault blocks that comprise the Pampean Ranges make up a region free of volcanic rocks, and conversely, the volcanic fields of the Argentina foreland are evidently not related to faulting.

El Valle Central is almost entirely in the batholithic zone, but prefers the eastern side at the north end.

PARANÁ BASIN BASALT FIELD

The Paraná basin is one of Paleozoic and Mesozoic age, developed by subsidence of a large, approximately oval-shaped region in the Precambrian Brazilian shield. The known Paleozoic section consisting of strata representing all periods except the Mississippian is at least 10,500 feet thick, The basin is about 1200 miles long and 400 miles wide (see map, Fig. 34.1). Desert conditions prevailed in mid- or early Late Triassic time and a windblown sand deposit was spread around irregularly. Then came the eruption of great floods of basalt. Between sheet flows in places more desert sand accumulated.

In southern Brazil these eruptive rocks are generally at least 400 m thick and are locally as much as 800 m. In Sao Paulo, north of Paraná and Rio Grande do Sul, the flows are locally separated by lenticular layers of cross-bedded eolian sandstone, some of which reach a thickness of 40 m. A characteristic of these extrusives is the general absence of olivine. Some lava flows are amygdaloidal, and these alternate with flows in which an irregularly developed columnar structure occurs. Pyroclastic rocks seem to be absent; the extrusion was of the calm type of fissure eruption.

Many feeding dikes and associated sills cut the underlying formations. Almost all the dikes are vertical. Most of the faults that cut the underlying formations also have steep dips. A number of fault planes, including some along which there was movement of 50 m or more, are filled by dikes of diabase. One of these, cut by the Santa Clara–Urubici highway on the top of the Serra do Panelao, Municipality of Bom Retiro, Santa Catarina, is a fault that vertically displaced the Botucatu sandstone (and apparently the basal part of the overlying eruptives) about 95 m. The fault is occupied by a diabase dike more than 300 m. thick (Avelino, 1956).

It may be calculated from the above figures and map extension of the field that about 75,000 cubic miles of basalt were extruded in fissure flows.

It is interesting to note that the Karroo system of sedimentary rocks in South Africa was invaded in Jurassic times by a multitude of diabase dikes and sills which crop out intermittently over an area of 1,500,000 square miles or about five times the area of the Paraná basalts. The volume has not been figured, but must be as much or more than that of the Paraná basin.

One half of the large island of Tasmania was once covered by diabase sheets which totaled at least 30,000 cubic miles. The Columbia River field contains about 40,000 cubic miles of basalt, also. All of these dike, sheet, and fissure-flow diabase and basalt fields are of the tholeiitic type.

Classification of Igneous Provinces

The Paraná basin field is clearly a tholeiitic basalt province, and it is evident that large volumes of primary tholeiitic basalt magma were generated and rose to the surface without differentiation. The origin of such a magma is a controversial question (Turner and Verhoogen, 1951), but it is generally agreed the source was below the silicic crust. The problem will be taken up later.

The voluminous volcanic rocks that accumulated prior to the batholithic intrusions with their abundant andesites and keratophyres are clearly of the eugeosynclinal andesite province, according to the writer's classification.

All the post-batholithic volcanic rocks that occur within the orogenic belt of the Andes are of the andesite orogenic belt province and are very similar to the eugeosynclinal rocks except that they lack the spilites and keratophyres. The rows of great stratovolcanoes which are very late in the general Cenozoic volcanic sequence are conspicuous for their alignment and dominant central vent character, but in terms of composition are must like the orogenic belt andesites with which they are closely associated. The melting of downward extended roots of mountains in the orogenic belts has been visualized as the source of the large volumes of andesite, but since large volumes of basalt, both olivine and tholeiitic, are also erupted in the orogenic belt with the andesite, we must provide for the rise into the granitic crust of such magmas from the subcrust. The basalt is generally more prominent in the late volcanic stages than as alternating extrusions with the andesite, and this fact should be kept in mind. Also it should be noted that by theory the roots of orogenic belts are thought to melt to form granodiorite and granite in great volume for the batholithic cycle, and on the other hand, some petrologists have postulated that roots melt in volume to provide the magma for the andesitic extrusions. Since the composition of granodiorite is considerably different, the same conditions, exactly, cannot exist for both.

35.

IGNEOUS AND TECTONIC PROVINCES IN MEXICO

GEOSYNCLINE

Very little is known of Mexico in Paleozoic time. In fact, it is not until Late Jurassic that much can be said of paleotectonic conditions when the Mexican geosyncline (Plate 10) began to form. It occupied central Mexico and extended longitudinally from Arizona to Mexico City (see Fig. 35.1). It is presumed to have been flanked on the north, west, and south by land areas, with the western land known as the Occidental geanticline. Up to 5000 feet of sediments accumulated in it, in large part an evaporite sequence. During Early Cretaceous time the geosyncline sank in places 12,000 feet to receive additional sediments of the miogeosynclinal type. Volcanism broke out on the west in Sonora with thick accumulations grading into the miogeosynclinal types on the east. The extent of volcanism is not well known, but altogether during Cretaceous time the deposits probably extended to the Pacific across what is now Baja California. (See Chapters 18 and 30.) Intense deformation of the geanticlinal area also occurred especially in the Early Cretaceous along the northern part, and coarse conglomerates were derived from the uplifted region, so we cannot characterize the area west of the miogeosyncline entirely as eugeosynclinal. Parts of it probably were eugeosynclinal, however, as indicated by the San Fernando formation of the northern part of Baja California. The eruptives are said to be andesite flows, tuffs, and agglomerates.

The extent of the volcanic area and geanticline is shown by the legend, pre-batholithic volcanic rocks, on the map of Fig. 35.1.

BATHOLITHIC BELT OF THE FIRST CYCLE

The Nevadan orogenic belt with its great granodioritic batholiths developed in the region of Baja California. This was the western margin of the eugeosynclinal and geanticlinal belt. The Lower and Middle Cretaceous sediments were folded and invaded by the batholiths and deeply eroded before the Upper Cretaceous sediments were deposited. The plutons are of immense size but have only been studied in northwestern Baja California, where they are typically quartz diorite. Reconnaissance reports generally refer to "granite." The metamorphic rocks have already been described in Chapter 30, and the belt may be summarized as typical of the Sierra Nevada in California and a continuation of it.

POST-BATHOLITHIC VOLCANISM

Minor disturbances and general uplift of Baja California, the Gulf of California, and adjacent Sonora followed, leaving a broad land area in this region. New volcanic outpourings occurred in the region of parallel ranges and valleys which are the foothills to the lofty escarpments of the Sierra Madre Occidental and in the Sierra Madre Occidental itself. These are the volcanic rocks that build the extensive Sierra

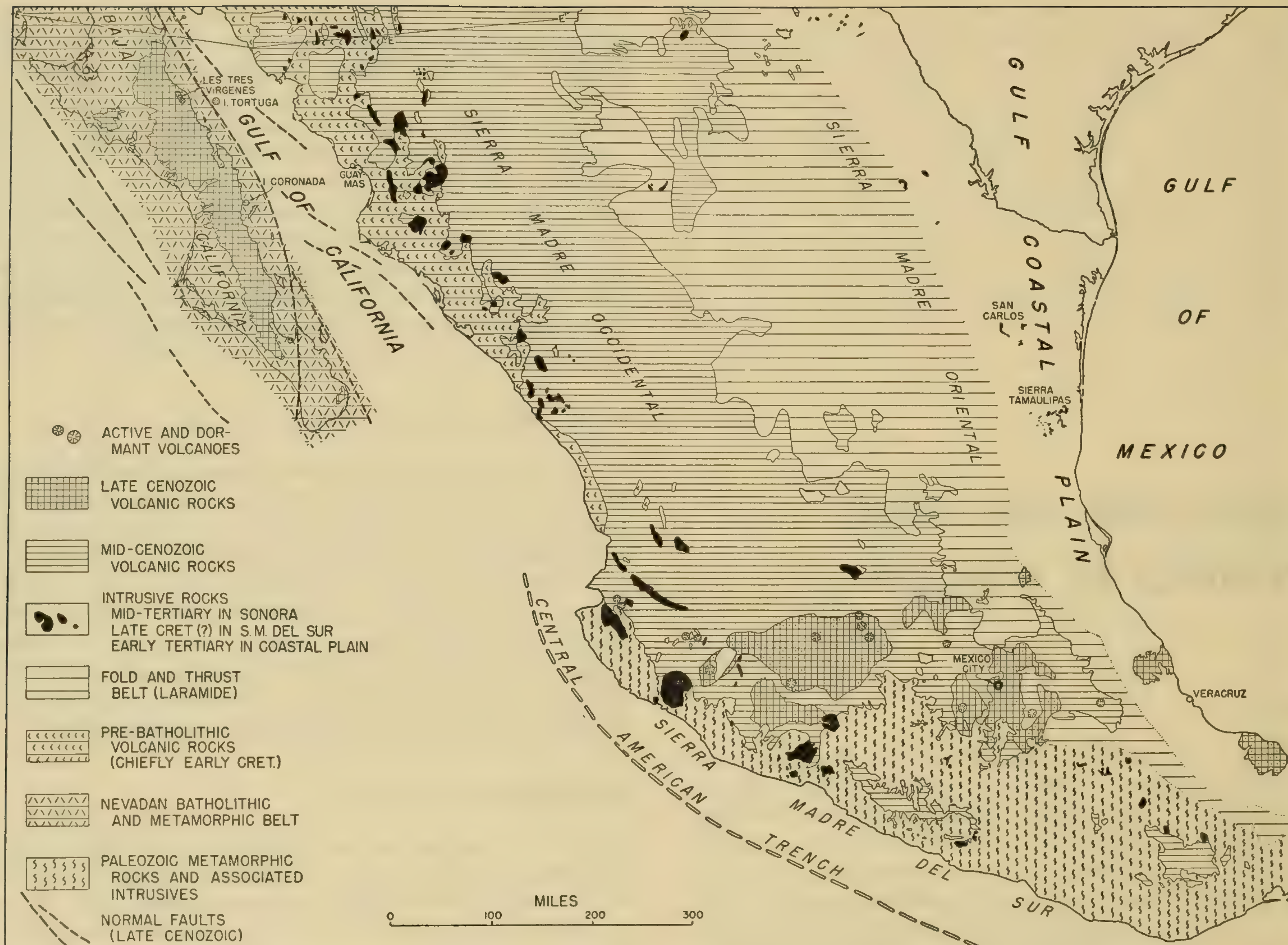


Fig. 35.1. Major tectonic and igneous units of Mexico. See Fig. 43.3 for active and dormant volcanoes.

Madre Occidental proper (see Fig. 34.5, section E–E′–E″). They are thought to be early Tertiary by King (1939) but the new geologic map of Mexico (1956) indicates them "principally as andesites of the Oligocene and rhyolites of the Miocene with corresponding pyroclastics." King also indicates that the Tertiary volcanic rocks are more acidic and more varied than the older beveled Cretaceous volcanic rocks upon which they rest in places in central Sonora. They contain a considerable thickness of rhyolite and some flows of basalt. In northeastern Sonora, Imlay (1939) notes that the lavas aggregate more than 5000 feet in thickness, and basalt predominates toward the top of the deposits but rhyolite and andesite are the most common. Basalt appears more abundant than in the region studied by King. Howell Williams (personal communication) recognizes large sheets of welded tuffs and thinks that these may be very extensive in the Sierra Madre Occidental. Much of the volcanic rock, heretofore called flows, at the north end of the Sierra Madre Occidental are welded rhyolitic tuffs (Enlows, 1955). The volcanic flows appear to be the result of fissure eruptions (King, 1939), but tuffs and pyroclastics indicate the occurrence of central vent eruptions also. The accumulations are thickest in the eastern Sierra Madre Occidental.

BATHOLITHIC BELT OF THE SECOND CYCLE

Along the western margin of the Sierra Madre Occidental, particularly in the region of parallel ranges of west-central Sonora, a mid-Tertiary (post-volcanic) orogeny occurred, and the volcanic and older rocks were folded in a measure exceeding the previous Laramide folding there. Accompanying the folding were vast intrusions of granite, diorite, and granodiorite which ascended through the Paleozoic and Mesozoic strata and in places penetrated the Oligocene and Miocene volcanic rocks. Granites predominate (King, 1939). These are the black areas on the map of Fig. 35.1 along the western margin of the Sierra Madre. In the Sonoran Desert geomorphic province the granites are carved to broad pediments, and the plutons are so extensive there that one may infer that the whole region is underlain by a vast batholith or series of large related plutons.

METAMORPHIC AND INTRUSIVE BELT

Extending across southern Mexico from Banderas Bay to the Isthmus of Tehuantepec is a belt of metamorphic rocks and various intrusive bodies. In width the belt extends from the coast to the Tertiary volcanic rocks of the Mesa Central which cover it irregularly on the north (Chapter 43).

Considerable parts of the belt shown on the map of Fig. 35.1 are covered with Jurassic and Cretaceous strata as well as fields of volcanic rocks whose age is not well known.

Although very little can be learned about the belt of orogeny, it seems evident that a pre-Jurassic and probable late Paleozoic age for most of it must be recognized. The Sierra Madre del Sur with its numerous post-metamorphic intrusions is regarded as a continuation of Baja California and therefore, of the Nevadan belt.

RELATION TO DEPRESSED BELTS

The Gulf of California is regarded as a depressed area along a zone of faults (Shepard, 1950). The faults in places have displacements comparable to those along the east side of the Sierra Nevada, and if the slope of their submarine escarpments has not been reduced since faulting, then the fault planes dip at rather low angles, which seems unusual. It is also observed that the San Andreas fault system extends through southern California to the head of the Gulf of California, and thence continues southward as the fault zone of the depressed Gulf area. Since the San Andreas and related fractures are generally recognized as a system of strike slip or wrench faults (Hill and Dibblee, 1953), a conflict in interpretation of the nature of faulting is evident. It is postulated in Fig. 31.22 that the block of Baja California has moved northwestward about 300 miles along the San Andreas fault zone and in so doing, has pulled away from the mainland somewhat, leaving the Gulf of California floored with oceanic crust. There can be no doubt, however, that the Gulf is a zone of subsidence in late Tertiary and Quaternary time. Anderson (1950) observed the faulting on islands in the Gulf to have extended from

Pliocene to Recent, and the zone is known to be one of modern seismic activity.

Along the adjacent western margin of Sonora, particularly in the province of parallel ranges and valleys, are fanglomerates with basal basalt flows and agglomerates of late Pliocene and perhaps younger age, the Baucarít formation (King, 1939). These have accumulated in downfaulted intermontane depressions. The basalt flows are generally conspicuous on the back slopes of tilted blocks where the overlying fanglomerates have been eroded away. The early or mid-Tertiary eruptives of the Sierra Madre Occidental are generally less basic.

Renewed orogenic activity resulted in overthrusting of rocks of each of the ranges west of the Sierra Madre westward over the Baucarít beds. This was observed north of the 28th parallel (Guaymas), but south of the parallel the faults are normal (King, 1939). In addition the western Sonoran normal and reverse faults are of the same age approximately as the Gulf faults and, hence, evidently belong to the same system. The reverse faults may be gravity slide phenomena. We have to deal, then, with a complex fault zone 150 miles wide in which submergence of the Gulf area relative to uplift of Baja California and the Sierra Madre was of the order of 10,000 feet.

Accompanying the faulting was the eruption of a large volcanic field on the southern part of the peninsula of Baja California. The accumulation is known as the Comondu formation which is made up of "many kinds" of volcanic rocks. The volcanism occurred possibly in Miocene time, but stratigraphically the flows seem related to the Baucarít formation of western Sonora of late Pliocene age. Comondu rocks may have been deposited near sea level and now are at elevations of 1000 to 5000 feet, which means adjustment of this order of magnitude along the great fault zone in Pleistocene time (See Chapter 30).

Here, in western Mexico, the downfaulted belt has developed along the continental side of the batholithic (Nevadan) belt, and evidently on the Cretaceous eugeosynclinal volcanic belt. This is a normal relation in reference to the Andean depressed belts. If the Paleozoic metamorphic belt exists here, it is mostly under the depressed area and covered. The second cycle batholithic belt is partly involved in the faulting, but mostly it is along the east margin of the fault zone. The great early and mid-Tertiary volcanic field of the Sierra Madre Occidental is east of the fault zone and suffered uplift at the time. Within the fault zone and on the west, on top of the first cycle batholithic belt, volcanism was recurrent. The field is of great extent in the southern part of the peninsula. Volcanoes were active during the Pleistocene and have continued active to the present. Isla Tortuga is a very young volcano in the Gulf, and Las Tres Virgenes are said to have been active in historic times. Isla Coronada is a Pleistocene andesitic volcano. Many cones and flows on the western slopes of Baja California exhibit features of recency (Beal, 1948).

The tectonic and petrologic relations of Baja California, the Gulf, and adjacent Sonora are similar to those of the Andes, but south of the Gulf the relations are less familiar. The Nevadan batholithic belt seems to be continued by the Paleozoic metamorphic belt and Mid-Cretaceous intrusions. North of the Paleozoic metamorphic belt is the southern termination of the great early and mid-Tertiary volcanic field of the Sierra Madre Occidental, and on top of these post-batholithic volcanoes and on the metamorphic rocks are great new piles of late Tertiary and Quaternary volcanics. According to Andean precedent these stratovolcanoes should be accompanied by a fault zone. The southern limit of the Mexican Plateau is said to be marked by a high fault scarp, but its position is not evident on the new geologic map of Mexico. The Balsas basin province may be due to downfaulting, but the writer has not been able to learn anything of the fault relations there.

36.

IGNEOUS PROVINCES IN WESTERN UNITED STATES

UGEOSYNCLINAL PROVINCE

The region west of the Antler orogenic belt in Nevada and California vas one of considerable volcanic activity in middle and late Paleozoic ime, especially in the Permian, and a thick assemblage of strata accumuated typical of the eugeosyncline. Volcanism persisted into the Mesozoic, nd in the mid-Triassic 12,000 feet of strata, chiefly pyroclastics and lavas, ccumulated to form the Excelsior formation. The rocks range in comosition from andesite through quartz latite to rhyolite with andesite robably predominating. Keratophyres with secondary albite have been dentified but probably have limited distribution. Certain intrusions which cut the lavas and tuffs of the Excelsior formation consist of much altered basic and also silicic porphyritic types. They are probably contemporaneous with the extrusions (Muller and Ferguson, 1936).

Following marine invasions and sharp folding and thrusting more thick volcanic deposits occur, which are of Jurassic age. These rocks were extruded during continued crustal unrest, and petrographically cannot be distinguished from the Triassic Excelsior volcanics. For further details refer to Chapter 6 and 17. Also examine map, Fig. 36.1 (symbol, prebatholithic volcanic rocks) for distribution.

BATHOLITHIC PROVINCE

Repeated Paleozoic, Triassic, and Jurassic orogeny occurred in the eugeosynclinal province before the deformed complex was invaded by the great batholiths. See Figs. 17.2 and 17.7. It has been pointed out that the Calaveras formation (Mississippian) is more metamorphosed than the Mariposa (Jurassic) in places, but it is clear that the Mariposa was sharply folded and thrust-faulted before the granodiorite intrusions. This has been regarded as a climactic orogeny immediately preceding the intrusions, but in the Sierra Madre del Sur of Mexico and in the South American Andes such an orogeny is either not evident or was of milder intensity, and the rocks into which the batholiths were emplaced are believed to be Paleozoic strata deformed and metamorphosed in late Paleozoic time.

The Sierra Nevada plutonic mass is a composite of many separate intrusions each of batholithic size. In the area of Yosemite National Park the individual batholiths made their ascent at about 2-million-year intervals over a period of 18 million years (Evernden *et al.*, 1957). The process of intrusion took place during Albian and Cenomanian time of the middle Cretaceous. Much of the rock is of forceful intrusive nature but considerable stoping, migmatization, and contamination of the primary magma occurred in places.

As shown in the cross section E–E′ of Fig. 34.1 the batholithic belt in central Baja California is about 175 miles (260 kilometers) wide, and in the California–Nevada region it has about the same width, if the satellitic

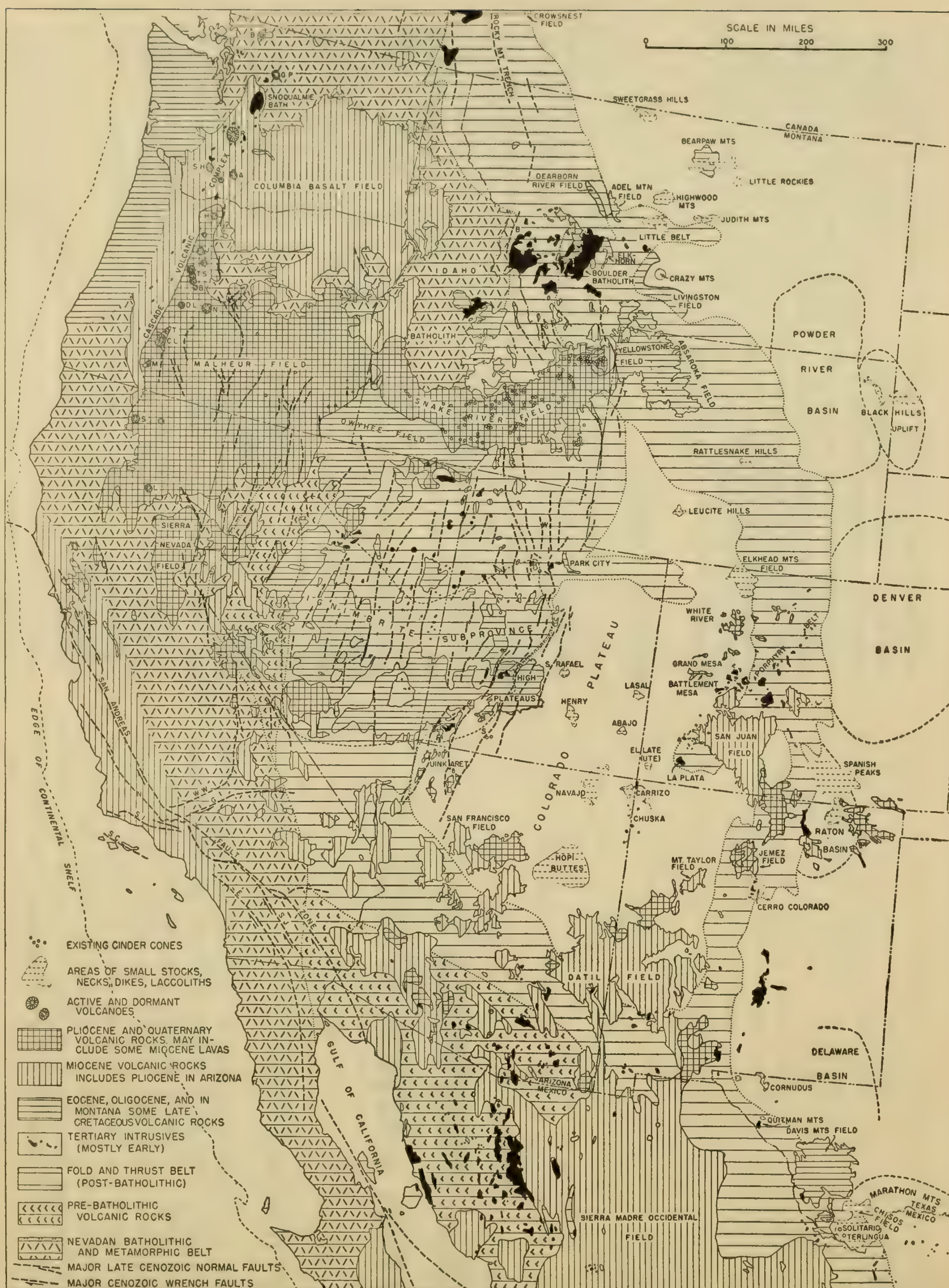


Fig. 36.1. Major tectonic and igneous units of the Western United States. L, Mt. Lassen; S, Mt. Shasta; M, Mt. McLaughlin; CL, Crater Lake; T, Mt. Thielsen; DL, Davis Lake; N, Newberry Crater; B, Belknap Crater; TS, Three Sisters; J, Mt. Jefferson; H, Mt. Hood; SH, Mt. St. Helens; A, Mt. Adams; R, Mt. Ranier; GP, Glacier Peak; B, Mt. Baker.

plutons in western Nevada are included. The pre-Franciscan metamorphosed sedimentary and igneous rocks exposed in the Coast Ranges of California seem to belong to a metamorphic belt such as was intruded by the batholiths in Chile, and may be west of the true batholithic belt.

In Oregon, Washington, Idaho, and southern British Columbia the belt is immensely wide—more so than at any other place. It has been pointed out in Chapter 17 that this region marks the intersection of two great arcuate segments of the Cordillera of western North America. The maximum width measured from the Cascade Range to the east side of the Idaho batholith is over 400 miles (650 kilometers). Farther north in southeastern Alaska and adjacent British Columbia it is about 300 miles wide, depending upon interpretations. By way of comparison, the Andean batholithic belt ranges from 40 to 70 miles in width.

In composition the great bulk of the Sierra Nevada batholith ranges from granodiorite to granite, with granodiorite indicated by some as the most voluminous, but quartz monzonite by others. See Chapter 17. Tonalite is said to be the dominant batholithic rock of southern California.

POST-BATHOLITHIC PROVINCES OF THE BATHOLITHIC BELT

Cascade Volcanic Complex

Divisions. The Cascade Range is a post-batholithic volcanic complex in Oregon and southern Washington (see map, Fig. 36.1), but in northern Washington and its continuation as the Coast Range of British Columbia it consists of the Nevadan complex. The central and southern volcanic part may be classed as an andesite orogenic belt province, and is divisible into the Western and the High Cascades.

Extrusive Rocks. According to Williams (1957) the Western Cascades:

> . . . consists of gently folded volcanic rocks ranging in age from late Eocene to late Miocene. Most of the topography here is mature and there are no traces of original volcanic forms. The High Cascades, on the other hand, consist of younger volcanic rocks that are virtually undeformed; most of the topography there is constructional and the original forms of the volcanoes, even though modified by glaciation, are easy to visualize. Other important contrasts distinguish the two belts. The thick volcanic accumulations of the Western Cascades are mainly products of fissure eruptions that produced extensive plateaus. Hence there are few eroded plugs marking the conduits of large volcanoes; instead, eruptive fissures are marked by narrow dikes of irregular trend. The High Cascades, on the contrary, were built almost wholly by eruptions from central craters so that clusters of large, coalescing cones were formed, many of which have been dissected by glaciers so as to reveal their feeding pipes. Finally, whereas the High Cascade volcanoes grew almost entirely by effusions of basalt and basaltic andesite, the rocks of the Western Cascades were produced by much more varied eruptions. Moreover these older rocks range in composition from rhyolite to basalt, and the lavas are intercalated with heterogeneous sheets of explosion debris, ranging from coarse agglomerates to fine tuffs, as well as with layers of tuffaceous sediment.
>
> The Western Cascade belt averages approximately 50 miles in width, and the volcanic rocks are as much as 13,000 feet thick. Beneath the High Cascades, these rocks must interfinger with equivalents of the Clarno, John Day, Columbia River, and Mascall formations, which are exposed on the plateau to the east.
>
> The High Cascade volcanoes probably began to erupt about the beginning of the Pliocene epoch, and almost all of them were broad shield volcanoes built by quiet outpourings of gray olivine basalt and subordinate flows of olivine-bearing basaltic andesite. Explosive activity contributed little to their growth until the final stages when the summit craters of many shields were capped by steeper cones of fragmental ejecta. Glacial erosion has modified the shapes of all these volcanoes; indeed, most of them have been reduced to radiating ridges separated by glacial cirques. The parasitic cones on their flanks have been all but demolished. The fragmental cones on their summits have been denuded until the more resistant fillings of their central pipes have been left standing as gigantic monoliths, like miniature Matterhorns.
>
> The earliest High Cascade lavas were erupted from a north-south chain of volcanoes close to the present edge of the Western Cascades. It seems moreover, that these volcanoes lay on or near the base of an eastward-facing erosion scarp cut in the rocks of the Western Cascade sequence. In places, this buried scarp was between 1,500 and 3,000 feet high, and where it was steepest and straightest it was almost certainly the result of faulting. As the volcanoes gained in height and the crest of the scarp was lowered by erosion, more and more of the High Cascade lavas were able to flow westward, inundating the scarp and spreading beyond on to a surface of low to moderate relief cut in the older volcanic rocks.
>
> The bulk of the High Cascades, as noted already, consists of Pliocene and Pleistocene olivine basalts and olivine-bearing basaltic andesites erupted from flattish shield volcanoes, and in places discharge of similar lavas continued until very recent times. But during the Pleistocene epoch several large, steep-sided, composite cones of andesite and dacite were built either on the tops of the older shields or in the depressions between them. The South Sister, for example is made up of three parts. Its lower part is an eroded basaltic shield volcano capped by a steeper cone composed of andesitic and dacite lavas, whereas its

upper part is composed of two Recent lava-scoaria cones of olivine basalt, the younger of which has a well-preserved crater that may have been active during the present millenium.

The largest Pleistocene andesite-dacite volcano was undoubtedly Mount Mazama, the ancestral mountain in the collapsed summit of which lies Crater Lake. This volcano, and its parasitic cone, Mount Scott, grew to full height by eruption of pyroxene andesites; then, in late Pleistocene time, more siliceous andesites and dacites were discharged from vents on a semicircular fissure on the northern slopes of the volcano, while a cluster of dacite domes rose near its eastern base and many basaltic cinder cones were formed elsewhere on the mountainsides.

During Pleistocene time, long flows of massive, pale-gray olivine basalt poured down the ancestral canyons of several of the principal rivers that now traverse the Western Cascades, such as the North Santiam, North Umpqua, and Rogue rivers, and the North Fork of the Willamette River. These flows did not issue from the central vents of the High Cascade volcanoes, but from fissures near the feet of these volcanoes and others farther west. They accumulated to a thickness of 1,600 feet in the ancestral canyon of the North Santiam, to about 1,000 feet in the North Umpqua, and to lesser thicknesses in other canyons. No doubt their eruption took place intermittently over a long span of time.

The principal eruptions of Pliocene and early Pleistocene time were from vents close to the crest of the range, but later eruptions are numerous on the eastern flank and on the adjacent plateau farther east. One of the most impressive recent lava fields is around and north of Belknap and Little Belknap crater. A line of cinder cones in the northern part of this field betrays rise of magma along a fissure. Another recent field stretches from Bachelor Butte through Sheridan Mountain to Lookout Mountain. More than 15 cinder cones and lava-scoria cones lie along a fissure system here.

A third large recent volcanic field is that around Newberry Crater (N, Fig. 36.1) which is 40 miles east of the crest of the High Cascades. According to Williams (1957) again:

The Newbury volcano is an approximately circular shield volcano about 20 miles in basal diameter which rises 4,000 feet above the surrounding plateau (Williams, 1935). On top there is a caldera, 5 miles long and 4 miles wide. The oldest visible lavas of the volcano are rhyolites exposed on the walls of the caldera. The rhyolites are overlain by basaltic flows and fragmental ejecta and by subordinate flows of andesite, and these in turn are capped by rhyolite flows that aggregate 1,000 feet in thickness, forming Paulina Peak. Presumably the volcano grew to its full height during the Pleistocene epoch; then its summit collapsed along ring fractures, probably in consequence of drainage of the underlying reservoir either by subterranean migration of magma or, more likely by copious eruptions of basalt from flank fissures. Thereafter eruptions took place within and outside the caldera. No basaltic flows and only a few basaltic cinder cones occur within the caldera, where most of the eruptions involved discharge of rhyolite. Outside the caldera on the flanks of the Newberry shield no less than 150 basaltic cinder cones were built and innumerable basaltic flows issued from them.

The row of stratovolcanoes of the High Cascades is continued northward by Glacier Peak (G. P.) and Mt. Baker (B) which are cones built on the Nevadan batholithic complex and isolated from the main volcanic complex of the Cascades. Even farther north in British Columbia 40 to 125 miles north of the city of Vancouver other Pleistocene volcanic cones occur. Mount Garibaldi (G on map, Fig. 37.1) has recently been described by Mathews (1958). There about 6 cubic miles of lava and pyroclastics have been erupted in good part during the Wisconsin stage of the Pleistocene. The extrusives are basalt and dacite; the dacite is most voluminous. Andesite in minor amounts is noted. Proceeding still farther north other volcanic mountains occur which are Mt. Clayley, Meager Mountain, and an unnamed one at 51°00′N. Lat.

These cones give the stratovolcanic row a length from Mount Shasta on the south to Meager Mountain on the north of 750 miles. The next known volcanic cone northward is Mt. Hoodoo, 400 miles from Meager Mountain, but it is possible that other volcanic cones occur between which have not yet been discovered. The rows of stratovolcanoes of the South American Andes range from 650 to 900 miles long, and hence are of the same order of magnitude as the Cascades volcanic row.

The volcanic rocks of the older Western Cascades are classed as tholeiitic by Waters (1955), and he describes them as pyroxene andesites and basaltic andesite constituting about 75 percent of the total and tholeiitic basalt, hypersthene basalt, and dacite pumice accounting for most of the rest. Some olivine basalt and rhyolte are also present.

Many of the lava flows and pyroclastic rocks contain abundant

xenoliths. Most are fragments of graywacke, silty argillites, greenstones, and basalts. Some show little change, others have been coarsely recrystallized and complexly modified by the enclosing magma. The abundance of inclusions in the andesites, and their near absence from the Eocene and Miocene basalts are noteworthy (Waters, 1955).

Much olivine basalt was erupted in the main growth of the Pliocene–early Pleistocene shield volcanoes and also in the late Pleistocene and Recent fissure eruptions. It is therefore evident that tholeiitic and olivine basalt kindreds are in close association and that magmas resulting from certain amounts of assimilation and subsequent fractional crystallization also played a role. After the volcanic rocks of the adjacent Coast Ranges of Oregon and Washington have been discussed, the origin of this complex suite will be considered.

Intrusive Rocks. Before leaving the extrusive rocks of the Cascade Range an intrusive group must be mentioned. According to Waters (1955):

> Numerous stocks and small batholiths of granodiorite and diorite cut the volcanic rocks. The largest is the Snoqualmie batholith, a composite mass of pyroxene quartz diorite, hornblende granodiorite, and granophyric quartz monzonite about 20 miles in diameter. The stocks occur in a linear belt along the core of the range [map, Fig. 36.1].
>
> Most of these intrusives are rather mafic augite–hornblende granodiorites and quartz diorites. Phenocrysts of plagioclase commonly show complex oscillatory zoning similar to that in the andesites. In places the Snoqualmie granodiorite is chilled against the enclosing andesites, but elsewhere the andesite is coarsely recrystallized at the contact and intimately penetrated by granodiorite. Miarolitic cavities are common. Parts of the granodiorite are altered; ferromagnesian minerals are decomposed to chlorite, the rock is cut by stringers of quartz and epidote, plagioclase is saussuritized, and albite, quartz, and epidote form irregular impregnations and replacements. These features suggest solidification under only a thin cover.
>
> The plutonic activity is not closely dated. The Snoqualmie batholith invades the Guye formation which contains fossil plants originally thought to be Miocene (Smith and Calkins, 1906) but now regarded as Eocene. The batholith had been deroofed by erosion before the building of the Mount Rainier stratovolcano whose basal lavas rest on grandiorites believed to be outliers of the Snoqualmie mass. The Shellrock Mountain instrusion of the Columbia River gorge cuts the Columbia River basalt and is overlain uncomformably by Quaternary andesites.

According to the classification proposed in this chapter, the stocks and batholiths of the Cascade Range are of the second cycle, whereas the batholiths of the Nevadan belt are of the first cycle.

Batholiths of two ages have recently been noted in the Vancouver area (Mount Garibaldi map area) by Mathews. The older underlies most of the area and is a heterogeneous assemblage of foliated and unfoliated quartz diorites and diorites. It is overlain unconformably by mid-Upper Cretaceous sedimentary rocks. The younger intrusive rocks consist of two plutons, one of which is a quartz diorite and trondhjemite and the other a quartz-rich granodiorite and quartz monzonite. Neither of the younger batholiths are in contact with the Upper Cretaceous beds, but they have escaped the extensive deformation which has tilted and block-faulted the stratified rocks, and are therefore considered younger than mid-Late Cretaceous. The potassium-argon age determination made by Follinsbee *et al.* (1957) appears to have come from the older batholith, for which an age of 105 m.y. is given. This is about Mid-Cretaceous and is consistent with the age indicated by the overlying mid-Upper Cretaceous beds.

The younger batholiths may correlate with the Snoqualmie batholiths of the Cascades of Washington, which according to Waters above, is post-Eocene and possibly as young as Miocene.

Coast Ranges Spilite and Keratophyre Province

Oregon–Washington Field. According to the classification of petrographic provinces proposed at the beginning of this chapter the Coast Range spilite and keratophyre province belongs to the eugeosynclinal class of "Andesite provinces." The western half of Oregon and Washington was a trough area of subsidence in which a great volume of volcanic rocks accumulated in Eocene and early Oligocene time (see Chapter 29). Weaver (1945b) estimates that more lava is represented here than the Columbia River basalt field, and Waters (1955) notes that more than 60,000 square miles were covered by the flows, and that in the northeastern Olympics the lavas are over 15,000 feet thick and in the Oregon Coast Ranges in a number of sections are over 6000 feet thick.

He estimates that the volume of Eocene basalt here is at least 40,000 cubic miles.

Petrographically the lavas are typical representatives of the tholeiitic magma type (Kennedy, 1933). They are aphanitic rocks composed of monoclinic pyroxene and labradorite set in a tachylitic base highly charged with magnetite dust. Phenocrysts of augite or plagioclase appear in some flows, but the series as a whole is characteristically nonporphyritic. Olivine is scarce or absent. Glass commonly accounts for 20–40 per cent of the rock. Chlorophaeite is abundant in some flows (Waters, 1955).

Waters points out that the flows at the bottom of a continuous sequence several thousand feet thick are of the same composition as those at the top or in the middle, and concludes that progressive differentiation had not occurred in the deep-seated magma chamber during the process of eruption. In contrast, the thick sills after emplacement show magmatic differentiation, and commonly consist of granophyric gabbro grading downward into feldspathic gabbro. The lower portions of the sills are rhythmically banded with layers of pyroxene and feldspar.

The basalts have been described in part as spilites, and the albitization in the Olympics has been pictured by Park (1944) as due to circulating heated sea water through the pillowed lavas on the sea floor. Waters (1955) does not reject this theory but believes it is not the entire explanation. He says:

Some dolerite sills, dikes, and subaerial flows are as thoroughly albitized as the pillowed flows. Albite veins and albite overgrowths on detrital feldspars are locally abundant in the graywackes and argillites that underlie the Olympic flows. Furthermore, most lavas might be better described as ordinary greenstones, zeolitized basalts, propylitized and saussuritized basalts, silicified basalts, and chloritized basalts, instead of spilites.

The Eocene basalts are underlain by thousands of feet of graywackes, argillites, and tuffaceous sediments. In the writer's opinion the alteration of the lavas to "spilites" and greenstones, and the simultaneous albitization, silicification, and chloritization of the underlying sediments and intrusive bodies have been produced by water, alkalis, silica, and other easily removable constituents stewed from the slowly metamorphosing root of geosynclinal sediments as it was downbuckled to form a tectogene. Fluids expelled from this metamorphosing root rose along zones of mechanical deformation altering the overlying volcanics and sedimentary rocks. This is essentially the same conclusion reached by Gilluly (1935) after an extensive review of the spilite-keratophyre problem.

California Field. Volcanic materials are observed in several of the Tertiary formations of the Coast Ranges of California but by all odds those of the Miocene are the most abundant, and are particularly well known in the central and southern Coast Ranges. The volcanic rocks are thickest in certain basins or around certain centers of volcanism, and in the central Coast Ranges several thousand feet of rhyolite tuffs, augite andesite, basalt, and olivine basalt flows occur in the San Luis Obispo–Huasna basin. Thick sills of analcite diabase and numerous plugs of andesite and rhyolite porphyries also occur.

In the southern end of the Santa Lucia Range there are rhyolite tuffs and flows and sills, flows of olivine basalt, often having a well-developed pillow structure, and numerous plugs of rhyolite porphyry. Rhyolite ash, basaltic pépérites, flows of basalt and numerous sills of analcite diabase occur in the Santa Cruz Mountains. Thin rhyolite ash, flows and breccias of basalt, and diabase sills are present in the Berkeley Hills, but they are not thick. Basalt flows occur in the Miocene of the Point Arena region. Aside from bentonized ash there are few volcanics in the Miocene in the San Joaquin Valley but there are numerous flows in the Cuyama Valley and the Carrizo Plain. There is abundant evidence that the volcanics were largely submarine; the tuffs and ashy sediments are often fossiliferous and the flows are generally interbedded with sediments containing marine fossils. It is possible that in some instances the volcanics accumulated so rapidly that local evanescent volcanic islands were built up, especially in the immediate vicinity of vents.

No single description would fit all of the occurrences of Miocene volcanics as the sequence and relative proportions of the various types vary somewhat. However the usual sequence is rhyolite tuffs and flows, flows of andesite and basalt, intrusions of sills and analcite and thomsonite diabase and intrusions of plugs, sills and dikes of soda rhyolite and waning explosive activity.

The sills of analcite diabase are an important and widespread phase of the Miocene volcanism. . . . Some of the thicker sills show gravitational differentiation and vary from a picrite at the base to a highly feldspathic diabase at the top. Most of them show chilled margins of analcite basalt, usually vesicular (Taliaferro, 1943b).

In the southern Coast Ranges 2280 feet of Miocene volcanic rock is exposed on San Miguel Island, 4700 feet on Santa Cruz Island, 8000 to 10,000 feet in the western Santa Monica Mountains and Conejo Hills, and at least 2000 feet in the area northeast of Glendora. Many wells have penetrated the same volcanics in the subsurface. Shelton (1954) estimates an average thickness of 1000 feet over an area of 700 square miles for

the volcanics of the southern Coast Ranges, and this would mean a volume of approximately 140 cubic miles.

Breccias and tuff breccias are most common but massive flows and intrusions are prominent in the Conejo Hills and Glendora areas. In the Conejo volcanic assemblage hornblende andesites occur at the base, and above these generally are breccias, tuffs, and flows of augite andesite. The upper part consists of flows of hypersthene basalt and olivine basalt. The basalts probably thicken southward in the subsurface. The intrusives in the area are chiefly diabase and hypersthene diabase (Shelton, 1954, 1955).

The Glendora volcanics are largely andesites, but olivine basalt and rhyolitic varieties are noted. In fourteen analyzed rocks the SiO_2 content ranges from 47.23 to 75.50 percent, and the most common types contain 59 to 63 percent. Present knowledge of the province as a whole indicates that andesites predominate among the extrusives, with basalt and dacitic or rhyolitic rocks following in that order. The associated intrusive rocks are dominantly basaltic or diabasic (Shelton, 1954).

Most of the volcanic rocks of this province are middle Miocene, but some may be slightly older. Shelton concludes that much of the lava was poured out on the sea floor or from vents close enough so that accumulation took place under water. Source fissures or vents have not been recognized. The relation of volcanism to tectonism is striking in the Los Angeles Basin. According to Shelton (1955):

> The Los Angeles basin is an area of locally derived Cenozoic sediments at least 25,000 feet thick, and as now exposed is a structural depression approximately 60 miles long and 40 miles wide. The most pronounced cycle in its history began in middle Miocene time and reached a climax of depth and localization during the upper Miocene and Pliocene. The climax of Miocene volcanism in southern California thus corresponds fairly closely with the beginning of the period of maximum growth of the basin.

Basalt Fields of Eastern Oregon and Washington

The Blue Mountains are composed of central island-like masses of Paleozoic and Mesozoic sedimentary rocks and intrusive masses with flanking volcanic flows and tuffs. See Fig. 29.15. The north flank volcanics are older and consist at the base of the Clarno formation of late Eocene (Duchesnean) and early Oligocene (Chadronian) age. It consists of a thick sequence of rhyolite and basalt flows with interlayered breccias and varicolored tuffs. Local unconformities are noted. See cross sections of Fig. 36.2. Overlying the Clarno is the John Day formation of late Oligocene and early Miocene age. It consists of colorful tuffs which in places may grade into acidic flows and breccias. Overlying the John Day is the Columbia River basalt which is now restricted to flows of mid-Miocene age. They are widespread in northern Oregon and southeastern Washington.

> The section at Picture Gorge along the John Day River [D, Fig. 36.2] may be considered as typical of this formation. Here it is situated between the John Day formation and the Mascall formation. The basalt series appears to be unconformable upon the John Day beds as shown by slight discordant relationships over a wide area, but appears to be generally conformable with the overlying Mascall formation.
>
> The Columbia River basalt poured out upon an area of varied relief. The basalt flows in places tend to be thick where they filled irregularities in the surface. The basalt flows are usually more massive and less columnar than flows high in the formation. Some flows contain appreciable amounts of olivine and weather more rapidly than the dense basalt higher in the section. Zeolites are particularly common in some of the basal flows, particularly in the Monument and Ritter quadrangles.
>
> The upper part of the Columbia River basalt characterized by "flow upon flow" structure is by far the thicker and more widespread part of the formation. Relatively parallel flows, commonly columnar, are visible for many miles along the canyon walls of northeastern Oregon. The upper flows are characteristically dark dense basalts with scoriaceous zones at the tops of each flow. According to Waters (1955, p. 708) continuous sections of more than 5,000 feet of basalt are found in northeastern Oregon (Baldwin, 1959).

Waters also calculates that about 35,000 cubic miles of basalt are present in the field.

The Mascall formation is largely made up of nearly white to buff bedded tuffs. It is late Miocene in age.

Following Mascall deposition the Columbia River basalts were folded and faulted near the Blue Mountains as shown in Fig. 29.15, and then eroded. On the erosion surface in mid-Pliocene time the Rattlesnake formation was spread. It consists of gravels, tuffs, and silts with a bed of welded tuff in the upper part. Uplift and moderate folding took place

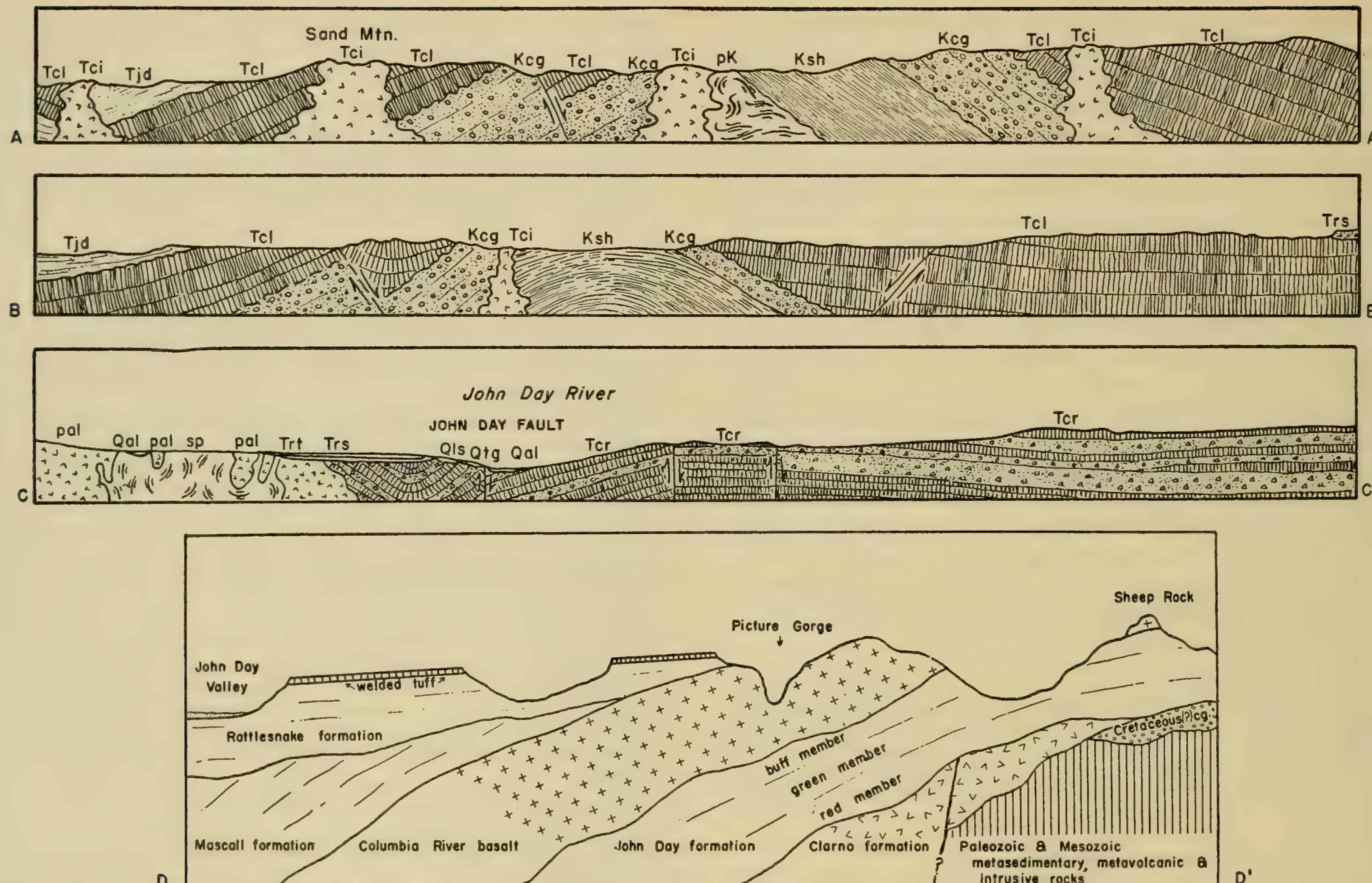


Fig. 36.2. Sections showing relations of Columbia River basalt to other Tertiary formations and to pre-Tertiary complex. A–A′ and B–B′ near Mitchell, Ore., on U. S. 26. C–C′ near John Day at junction of U. S. 26 and 395. Reproduced from Wilkinson, 1959. D–D′ is schematic of Picture Gorge area, John Day River. Reproduced from Baldwin, 1959.

pal, metavolcanic and sedimentary rocks; sp, serpentine; pk, pre-Cretaceous rocks; Ksh, Cretaceous shale; Kcg, Cretaceous conglomerate; Tcl, Clarino lavas; Tci, Clarno intrusives; Tjd, John Day formation; Tcr, Columbia River basalt; Trs, Rattlesnake fm.; Trt, welded tuff.

after the deposition of the Rattlesnake formation. The Pleistocene in central Oregon was mostly a time of erosion.

The Blue Mountains are flanked on the south by Mio-Pliocene volcanics of the Payette and Owyhee formations and correlative beds. The pre-Columbia River basalt formations are missing along the southeast side of the Blue Mountains and the Payette, oldest in the area, is correlated with the late Miocene Mascall on the north side. The High Lava Plains (Fig. 29.15) south of the Blue Mountains are made up of relatively undeformed young lava flows dotted in places by cinder cones and lava buttes. The formations are dominantly Pliocene lavas, tuffs, and alluvium, few of which have been formally named (Baldwin, 1959).

Basalt Kindreds. H. A. Powers of the U.S. Geological Survey has commented in a letter to the writer about the problem of basalt kindreds in the northwestern states, and has charted the chemical analyses of about 65 characteristic basalts in regard to SiO_2 and MgO from the Columbia Plateau, the Snake River downwarp, the Malheur Plateau and Hawaii. He finds such a scatter of points that the concept of a clear-cut distinction of tholeiitic and olivine basalt seems to break down. The Columbia River basalts of Miocene age run relatively high in SiO_2 and low in MgO; the Hawaiian basalts classed as tholeiitic run slightly less in SiO_2 and intermediate in MgO; Hawaiian rocks classed as olivine basalts are intermediate to low in SiO_2 and low, intermediate and high in MgO; the Snake River Pliocene and Recent basalts run generally low in SiO_2 and intermediate in MgO; the Steens Mountain basalts in the Malheur field run intermediate to low in SiO_2 and generally low in MgO. As a result he says:

> In some provinces, there is a decided gap, or absence of rocks showing all the intermediate stages. In such provinces there appears to be an impressive difference between tholeiite and olivine basalt, in the chemical sense. My feeling is that the concept of a fundamental distinction between two kindreds of basalts has been developed from a concentration on such single provinces, but that the concept breaks down and is not convincing when one considers all the basalts that we know about from good comparable chemical analyses. I have plotted in different ways about a thousand reasonably good analyses of basalts trying to establish a natural division zone, and so far have succeeded only in showing a complete gradation—a lot of crossbreeding if there are really two kindred.

On the other hand, he believes that perhaps a difference can be made between flood eruptions and cinder cone or small lava dome eruptions, and that this may reflect fundamental differences in the tectonic setting. Such a distinction is based on the field characteristics and not on the chemical compositions. In the Columbia River basalt field flood basalts predominate and are presumed to have issued from fissures. Most of the Pleistocene basalts in the Columbian River field are fissure flows also, but some seem to be of cinder cone activity (Powers, personal communication). The Snake River and Malheur fields, on the other hand, are mostly of the cinder cone and small lava dome type.

Snake River Basalt Field. The eastern part of the Snake River lava plain from King Hill and Twin Falls to Yellowstone Park, a distance of 200 miles, has been studied in considerable detail by Stearns, *et al.*, (1938). They report that about 95 percent of the rock of the depression or downwarp is the so-called Snake River basalt of Pliocene, Pleistocene, and Recent age. Locally sedimentary lenses, closely related petrologically to the flows, exist, and some of these are very fossiliferous such as the Hagerman lake beds. In numerous places on the borders of the plain rhyolitic flows and pyroclastics emerge from beneath the basalts. Perhaps the rhyolites are younger and stratigraphically above the Challis volcanics on the north border which are dominantly latite and andesite. The Challis volcanics are regarded from fossil leaf beds as late Oligocene or early Miocene, and ages up to early Pliocene have been assigned to the rhyolites. At places rhyolites crop out within the basalt plain under the basalt, and hence it is believed that the rhyolite volcanics extend widely under the field and form the basal layer (Kirkham, 1931).

The rhyolites have been loosely referred to as the Mount Bennett rhyolite and Owyhee rhyolite, but much of the rock is quartz latite or even possibly andesite similar to the Challis volcanics (Stearns *et al.*, 1938).

Three old cones are prominent landmarks in the area between Arco and Blackfoot, and their building seems to predate the Snake River basalt. Big Southern Butte, about 5 miles in diameter, rises nearly 2500 feet above the plain and is composed of basaltic and rhyolitic flows. The main mass is a light-colored porphyritic rock containing large quartz

crystals, and has been identified megascopically as rhyolite. The cone is much eroded.

East Butte is made up of beds of trachyte, pumice, and obsidian, which strike east-west and dip 30 degrees south. No vestige of a crater remains, and it is possible that the butte is part of a tilted fault block. The third butte, known as West or Middle Butte, lies 4 miles away. It is composed entirely of basalt which dips 10 degrees south. If East and West Buttes are both parts of the same tilted fault block, then interlayered trachyte and basalt must be postulated. Whether a fault block or separate cones, they were deeply dissected by erosion before the Snake River basalts were spread around them. A thin section of the basalt of West Butte shows "abundant feldspar, olivine, and pyroxene, with a little brown glass."

A number of units in the Snake River plain are younger than the rhyolites yet older than the basalts that cover most of the plain. They are mostly basalts and associated lake beds. The extensive Pleistocene and Recent basalts are said by Stearns to have come from about 400 vents in the plain. He charted the position of about 300 of them. Except for the cluster in the Craters of the Moon National Monument and the group north of St. Anthony, they are rather evenly distributed and neither a rift nor fault pattern is discernible, although here and there short rows of cones occur.

Near the north side of the Snake River Plain cinder cones 50 to 200 feet high predominate. However, over most of the plain the vents are broad lava domes each usually about 100 feet high and the related flows covering about 30 square miles. Only a suggestion of a crater or crater rim is left generally when eruption ceases. The lava welled out quietly and profusely and each vent had only one period of activity. With activity over in one vent another one nearby seems to have formed and poured out considerable lava.

The geology of the western part of the Snake River volcanic field has been summarized by Kirkham (1931). He believes that the basal layer is a Miocene basalt and that this is very widespread. He calls it the Columbia River basalt, but describes it principally as an olivine basalt which does not correlate with the tholeiitic basalts of the Columbia River basalt field proper. This basal unit has been eroded irregularly and its existing thickness in outcrop ranges from 300 feet to over 1200 feet. The basal "Columbia River basalt" occurs in three stratigraphic parts, namely, lower and upper basalt flow units and intermediate lake beds containing much tuff, the Payette formation.

The Owyhee rhyolite, previously mentioned, rests on the basalt, at least in the area of southwestern Idaho south of the Snake River. Kirkham states that the rhyolite is actually a series, and is generally made up of basalt and andesite flows at the bottom, and above by trachyte, latite, and rhyolite flows interbedded with ash, fresh-water limestone, clay, shale sandstone, and conglomerate layers. He correlates the series with the Salt Lake formation south of the Snake River plain. The distribution and stratigraphic and petrographic relations of the "Columbia River basalt" and Owyhee "rhyolite" seem to need much more study before the picture can be significantly summarized.

Above the Owyhee rhyolite and Salt Lake beds is the widespread Snake River basalt, so characteristic of the eastern part of the field previously described. The Snake River basalt flows give way to and are covered by lake beds in western Idaho which are known as the Idaho formation (Kirkham, 1931), but here as in the eastern part of the plain, the Quaternary history was eventful with repeated, if scattered, constructional volcanic activity, struggling against the destructional activity of the Snake River for supremacy (Norman Anderson, personal communication).

The Snake River volcanic field together with the Malheur and Columbia field constitute a unique petrographic province from the tectonic point of view. The western part of this great field covers the Nevadan batholithic and orogenic complex, and the eastern arm lies across the Laramide fold and thrust belt of the central Rockies (Chapter 22). We are accustomed to a parallel arrangement of volcanic deposits with the orogenic belt; even if discontinuous in extent, the volcanic fields do not take a transcurrent direction. Here, however, the eastern arm of the Snake River field extends almost at right angles over the underlying folds and thrust sheets of southeastern Idaho and southwestern Montana.

Malde (1959) reports a great fault zone along the northern boundary of the Snake River Plain in the area west of Boise. Gravity, seismic, and geologic studies indicate that at least 9000 feet of aggregate throw has

displaced the Plain downward relative to the highlands on the north. At least 5000 feet of movement occurred between the early and middle Pliocene, and progressively diminishing movements amounting to 4000 have occurred since.

The crustal break implied by the gravity measurements is possibly expressed by a line of earthquake epicenters that extends diagonally from Puget Sound, across the Columbia River Plateau, along the northern boundary of the western Snake River Plain, and thence across the plain to northern Utah. In Idaho, these earthquakes originate principally at average depths of 61 and 38 km (38 and 24 mi), the shallower earthquakes being near the base of the crust (6). The displacement calculated from the gravity measurements therefore ranges from one-tenth to one-third of the local crustal thickness (Malde, 1959).

The geology of northern Utah hardly permits the extension of the fault zone into this region. The writer believes, rather that a more logical projection is eastward under the Snake River volcanic field to Yellowstone Park. It is thus shown on Figs. 31.21 and 31.22, where its tectonic significance is discussed. It is interpreted chiefly as a zone of distention, and if so, seems to afford a natural channelway for the lavas from the base of the silicic crust and from the basaltic subcrust. See Fig. 31.25. The transcurrent nature of the Snake River volcanic field is thus better understood. Also, the fissure effusion of great volumes of basalt from the subcrust may be accounted for.

PROVINCES OF THE MIOGEOSYNCLINE AND SHELF

General Characteristics

The tectonic provinces of the Rocky Mountains stand apart from the Pacific marginal provinces in several respects; their mountains, plateaus, and basins were developed by late Mesozoic and Tertiary orogeny and epeirogeny in the Paleozoic miogeosynclinal and shelf regions and also on the miogeosynclinal-type sediments of various Jurassic, Cretaceous, and Tertiary basins. Thick late Precambrian sandstone and shale sequences underlie part of the Paleozoic miogeosyncline and shelf areas, and in other areas, particularly in Colorado and Arizona only a very thin sedimentary veneer existed on the crystalline rocks of the Precambrian basement at the time of late Cretaceous and Tertiary orogeny. Such is the general tectonic setting for the eventful and diversified igneous history of the Rocky Mountains which began in Cretaceous time and continued from place to place to the present.

The igneous rocks of the Rocky Mountains, like the sedimentary rocks and structures, stand apart fairly distinctly from those of the eugeosynclinal and batholithic belt to the west; in particular they are generally more alkalic. Basalts and andesites are present and in places abundant, and the orogenic type basalt-andesite-dacite-rhyolite association is prominent, and therefore a similarity exists with this overwhelmingly preponderant kindred of extrusive types in the Pacific marginal regions. But where present the intermediate latitic differentiates are most abundant in contrast to the dominant andesites of the Pacific marginal belts. The Rocky Mountains are characterized especially by the classical kindreds of calc-alkalic olivine basalt-trachyte-phonolite and alkalic leucite basalt-trachybasalt-trachyte. The nepheline syenites are intrusive accompaniments in places. For the fractional crystallization associations an olivine basalt is generally considered the parent magma, but assimilation or fusion of small or appreciable amounts of calcic or alkalic country rock such as limestone, amphibolite, granite, or mica schist by the olivine basalt magma is postulated, or at least admitted as possible, to produce the melts from which the high calc-alkalic or alkalic fractional crystallization kindred resulted.

Trachyte and Phonolite Provinces

Extent of Provinces. Igneous rocks containing a high amount of either sodium or potassium or both are characteristic of large areas in the Rocky Mountains.

Three high alkalic kindreds are generally recognized on a world-wide basis, the leucite basalt-trachyte, the olivine basalt-phonolite, and the nepheline syenite (Turner and Verhoogen, 1951). The first two are classed as nonorogenic assemblages and the last, which is, of course, an intrusive type is regarded as a low-temperature, high alkalic residue of an evolutionary series in which volatiles played an important role. The phonolites, trachytes, and syenites appear as minor end members of an olivine basalt

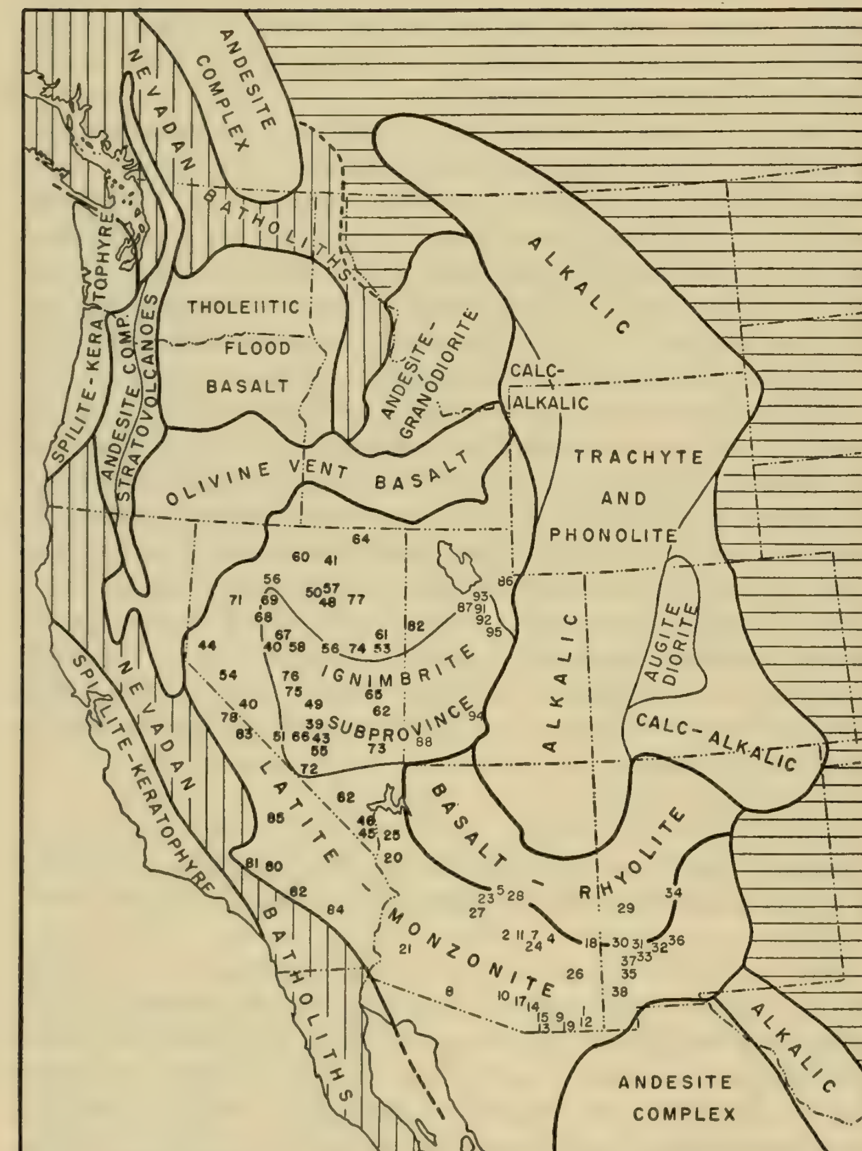


Fig. 36.3. Igneous provinces of the western United States. The numbers relate to intrusions listed in the table on page 574.

parentage. The writer has not found it possible to chart these three kindreds in separate provinces in the Rocky Mountains, and therefore does not try to distinguish them. They will be referred to collectively as the trachyte and phonolite province. (see map, Fig. 36.3). Igneous rocks adjacent to the region of alkali-rich igneous rocks in the Rocky Mountains are generally more calcic or do not display an excess of alkaline elements such as to yield the feldspathoid minerals, and are grouped in calc-alkalic subprovinces.

Colorado Plateau. The chief igneous centers in the Colorado Plateau which belong to the high alkalic subprovince are the laccolithic groups (Henry, La Sal, Abajo, Ute, and Carrizo Mountains), the Navajo and Hopi Buttes volcanic fields of northeastern Arizona; and the San Rafael Swell. The Elkhead Mountains, White River, Grand Mesa, and Battlement Mesa fields are also of alkalic affinities and are grouped in the Colorado Plateau for convenience sake.

In the laccolithic groups (Hunt, 1954, 1956) the first intrusions are diorite porphyry which constitutes about 60 percent of the total volume of igneous rock. Intrusions of monzonite porphyry follow to the extent of about 25 percent, and then last a syenite porphyry to the extent of about 13 percent. The last intrusion is noted only in the La Sal Mountains. The rocks are high in Na_2O, but the ratio of K_2O to $CaO + Na_2O$ increases eastward. The earliest intrusions in each group contain about 5 to 6 cubic miles of rock. These were stiff and relatively low-temperature magmas. The central stocks probably breached the surface and erupted more potassic rock than contained in the intrusions.

The magmas were intruded in basins, broad domes, and benches of the Colorado Plateau. Olivine basalt is regarded as the primary magma which assimilated amphibolite and hornblende gneiss to yield a potash-rich magma which then differentiated (Waters, 1955).

The Hopi Buttes (Williams, 1936) is a volcanic field of lava-capped mesas and many necks. Ejecta consists of limburgite (dark, glass-rich and usually minus feldspar) and monchiquite (nepheline basalt) in sedi-

mentary matrix. Lavas are analcite basalt. Feldspar is scarce or absent and analcite abundant. MgO, CaO, and Na_2O are high; K_2O is low.

The Navajo volcanic field (Williams, 1936) consists principally of a number of necks of tuff breccia and agglomerate crowded with fragments of granitic rocks. These breccias and agglomerates are high in K_2O in contrast to the Hopi Buttes rocks, and fairly low in Na_2O and fairly high in MgO and CaO, and have been called sanadine-rich trachybasalts and leucite basalts. Williams suggests that an originally sodic ultrabasic magma having the composition of nepheline basalt reacted with the potash feldspar of granites in the basement and so attained the high potassic composition which prevails in the subprovince.

In the interior of the Plateau, in the laccolithic mountains, soda greatly exceeds potash. The same is true in the Hopi Buttes field along the southern edge of the Colorado Plateau, but in the intervening Navajo field potash greatly exceeds soda.

The Elkhead Mountains of northwestern Colorado constitute a high alkalic volcanic field. The suite is unusual with rocks containing both olivine and quartz, a nepheline-bearing trachyte with phenocrysts of yellow-brown mica in a groundmass of sanadine and nepheline, and analcite basalt without feldspar, and with dikes of soda verite, analcite syenite, and soda syenite (Carey, 1955).

Central Wyoming. Leucite Hills are located in south-central Wyoming on the north end of the Rock Springs uplift. They are remnants of lava flows and cinder cones on a mid-Tertiary erosion surface, now much dissected and left about 800 feet above the present valley floors. The rock is called Wyomingite, and contains phlogopite, leucite, and diopside (Cross, 1897).

The Rattlesnake Hills field of central Wyoming consists of three large necks and a number of small necks and related dikes in an area of 150 square miles. The first and largest intrusions and extrusions were viscous, acid quartz latites. Following these a series of highly alkalic trachytes, phonolites, and vogesites were erupted. (Vogesites are lamprophyres, generally considered to be hypabyssal.) The alkalic rocks are unique for their content of the relatively rare feldspathoidal minerals, huayne, and nosean. Although the necks are in a rather small area, the amount of material ejected was large and certain clastic parts are believed to have been transported 100 miles from the volcanic center. The activity is dated as mid-Eocene (Carey, 1954). Most of the immediate ejecta has since been eroded away, but water-transported fragments are prominent in a middle and upper Eocene formation of the general region (Van Houten, 1955).

Black Hills. Across the north end of the Black Hills uplift is a row of imposing Tertiary volcanic necks and laccoliths in Mesozoic strata known from west to east as Devils Tower, Bear Lodge Mountain, Bear Butte, Inyankara Mountain, and Mineral Hill. These are composed of phonolite, pseudoleucite porphyry, nepheline syenite, and aegerite syenite (Robinson, 1956).

Several of the centers of Tertiary igneous activity are domal uplifts in the Paleozoic and Mesozoic sedimentary rocks and the underlying cause of doming is regarded by Noble *et al.* (1949) as due to the intrusion of stocks rather than laccoliths. One of the domes includes the noted Homestake gold mining district at Lead. It is 10 by 12 miles in size and contains several rather ragged Tertiary stocks and numerous sills and dikes. The intrusive rocks have been described as phonolite porphyry, rhyolite, and quartz porphyry (O'Harra, 1933).

The entire domal structure of the Black Hills, some 50 miles by 100 miles, is considered possibly due to a major Tertiary batholithic intrusion by Noble *et al.*, but they see no way of finding evidence of the intrusion. The gravity picture which might help is clouded by the dominance of gravity lows over the adjacent Cretaceous and Tertiary basins.

Central Montana. North-central Montana is characterized by a number of mountain groups, each of which owes its existence to igneous activity, both intrusive and extrusive. The region is east of the Laramide belt of intense compression and the magmas have penetrated nearly horizontal sedimentary strata.

The rocks range from rhyolites to basalts in one category and from shonkinites through nepheline syenites to syenites in another. The rocks of the latter category are rich in potash and soda and almost devoid of plagioclase. The rocks of each mountain group fall into one or more eruptive stages; and the rocks of each stage have peculiar mineral and

chemical features, although they commonly range from highly mafic to highly felsic. Each stage is separated from the other by intervals during which few or no eruptions occurred, but instead, extensive erosion.

In each of the stages a rock near the mafic end is believed to represent the primary magma. This rock ranges from an ordinary basalt to orthoclase basalt to plagioclase shonkinite to shonkinite rich in potash and lacking plagioclase. The gradational character of the eruptive stages and their close association in time and space indicate a common origin. Two periods of magmatic differentiation are required: first, a deep-seated differentiation of a basaltic magma from which crystals of calcic plagioclase and hypersthene were removed and second, a shallower differentiation to form the magmas of the individual eruptive stages. The relative flatness of the sedimentary rocks into which and through which the magmas have moved indicates that the magmas have not been disturbed by orogenic forces; therefore they could have differentiated during the long quiet interval which seems necessary. The second period of magmatic differentiation by crystal settling was characterized, in most stages, by assimilation of siliceous material. The amount of assimilated material was especially large in the Crazy and Little Belt mountains, where syenites were followed by granites (Larsen, 1940).

The abundant flows and dikes of mafic phonolite, and flat laccoliths and dikes of chemically equivalent shonkinite are derivatives of basic potassic magmas. Syenite is undoubtedly a differentiate of a parent shonkinite magma after intrusion as a sill or laccolith (Turner and Verhoogen, 1951). Larsen (1940) believes essentially that all petrographic and chemical variations within this region may be explained in terms of magmatic differentiation from an olivine basalt. A long perod of undisturbed differentiation in depth is required in which settling of olivine and diopsidic augite takes place to leave the melt enriched in K_2O. Turner and Verhoogen (1951) would place more emphasis on reactive assimilation with the granitic basement.

Summary. The province of high alkalic rocks has the following characteristics:

1. The region is one of crustal stability for the most part. It was a shelf to the west-lying miogeosyncline and part of the interior stable region in Paleozoic time. Triassic and Jurassic deposition was thin but Cretaceous sediments accumulated in several separate intermontane basins to a thickness of about 5000 feet. The total section of nearly flat-lying sedimentary rocks did not exceed 10,000 feet in any place, and in some areas, as in central Colorado, only a few hundred feet of sedimentary rocks existed at the time of igneous activity.

2. The relatively thin veneer of sedimentary rocks rest directly on metamorphosed crystalline rocks, generally of a gneissic or schistose character. In the region of high alkalic rocks no Beltian type rocks are known, except in west-central Montana on the border of the alkalic province. This feature correlates well with the common observation of granitic, gneissic, and amphibolitic inclusions in rocks of a number of the igneous centers, and also with the conclusion that such crystalline rocks have been assimilated in various amounts by an olivine basalt magma. The inference is warranted that olivine basalt underlies the "granitic" crust directly, that the primary activity begins in the basaltic layer or subcrust, then proceeds to the granite crust where assimilation takes place. With stable crustal conditions prevailing, the various alkalic rocks originate through fractional crystallization, intrusion, and further differentiation.

3. This is a region of high BaO and SrO and also of the most abundant uranium ores so far discovered in the West. Such elements may have been derived from the assimilated Precambrian crystalline rocks and later concentrated by differentiation. The U_3O_8 would be further concentrated by meteoric or epithermal processes.

4. No basalt is found in the laccolithic groups, but these igneous centers stand apart from the others in having only small volumes of intruded magma and relatively stiff cold magmas at the time of intrusion. In the other fields, in fact in most all volcanic fields of any size, basalt is erupted generally either early or late in the history of the field, and therefore we must think of a facility whereby some basalt from the subcrust makes its way directly to the surface without an intermediate rest stage for assimilation or differentiation.

5. The Rockies of Montana, Wyoming, Colorado, New Mexico, and Utah including the Colorado Plateau, are east of the fold belt of the central Rockies and are the result primarily of large domal uplifts with lateral gravity slide affects in places. See Fig. 25.12. The surficial igneous centers in the trachyte-phonolite province occur in the basins, domes, and across monoclinal flexures, and graben. If the domal uplifts are supported

by downward protuberances of the granitic crust or of the basaltic subcrust, and if these are melted and responsible for the location of the igneous centers, such as is generally held to be the case in the Nevadan orogenic belt, then the upward coursing magma must have worked laterally considerable distances to have found outlet in the interuplift sedimentary basins.

The domal uplifts are structures caused by vertical forces, and hence it is believed that roots could not have developed; roots are the result of horizontal compression or crustal shortening. The conclusion seems evident that the domes are themselves the result of igneous activity; they are great blisters above giant laccoliths or thick megasills in the "granitic" layer. The original magma in the megasills is postulated to be olivine basalt, which while still molten, assimilated variable amounts of the crystalline basement, and then as a secondary magma intruded through the overlying crystalline basement and the sedimentary veneer to the surface. In certain places like the Henry Mountains, minor amounts worked somewhat laterally to emerge in the adjacent basin. The position of some of the igneous rocks which have penetrated the sedimentary veneer poses a problem, it must be admitted, but then, to the writer's knowledge no attempt has been made to explain their distribution by any other hypothesis.

The blister concept is illustrated in Fig. 36.4.

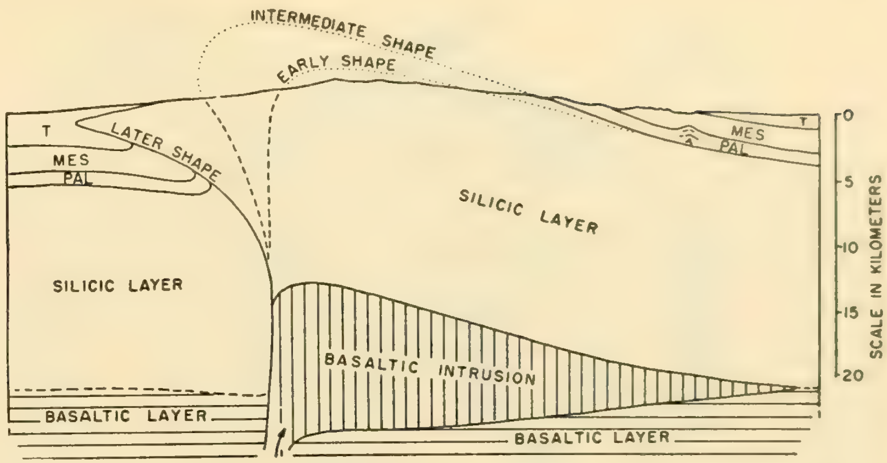


Fig. 36.4. Concept of blister structure and gravity mass movements of oval-shaped uplifts in shelf province.

West Texas and Mexican Coastal Plain. The principal volcanic field in the west Texas province is the Davis Mountains which extend from the southern flank of the Delaware basin to and across the Rio Grande into Mexico, a distance of 125 miles (*Tectonic Map of the United States,* 1944). The Chisos Mountains and the Terlingua–Solitario region to the southeast in the Big Bend Country, have many igneous bodies. A number of intrusives are known in adjacent Mexico in the Sierra Madre Oriental and Serrania del Burro uplift. Northwest of the Davis Mountains are the Eagle Mountains and Quitman Mountains which contain intrusive and extrusive bodies, and north of these and east of El Paso are a group of small intrusives that make up the Cornudus field. The Marathon basin also contains a number of plugs and dikes.

An alkalic composition has been noted in many of the igneous rocks of west Texas but some are calc-alkalic. The overall province, however, is generally referred to as alkalic and related to the Spanish Peaks and central Montana provinces.

The extrusive rocks of the Davis Mountains are trachytes, phonolites and some rhyolites. Intrusive rocks are syenite and sodic syenite porphyries. Olivine basalt occurs in minor amounts. All these igneous rocks are early Tertiary in age, but one Recent vent has been observed (King, 1937).

The igneous rocks of the Cornudus field are augite syenites and analcime nepheline syenite.

A volcanic area in the Quitman Mountains has a ring-dike and stock of quartz monzonite as a central feature. This locally cuts a volcanic series which consists of lower rhyolites, intermediate trachytes, rhyolites, latites, and andesites, and upper trachytes. The total thickness is about 3500 feet, and rhyolite appears to occur in largest amounts. Indirect fossil evidence suggests an early Tertiary age. According to the alkali-lime index of Peacock, the volcanics of the Quitman Mountains fall near the boundary of the two intermediate series, alkalic-calcic and calc-alkalic (Huffington, 1943).

Fig. 36.5. The Solitario. Simplified from E. H. Sellards, W. S. Adkins, and M. B. Arick. Unpublished map from Bureau of Economic Geology, University of Texas.

The Terlingua–Solitario region is one of profuse and diversified igneous rocks. According to Lonsdale (1940), there are several hundred masses distinct enough to be mapped in an area of about 400 square miles. They occur as lava flows, plugs or necks, dikes, sills, laccoliths, bysmaliths, and possibly stocks. The largest plutons are laccoliths. Solitario is the largest domed-shaped structure of the group and is strikingly circular. It may be a laccolithic dome (see Fig. 36.5). The igneous rocks of the district include an analcite-bearing series which ranges from melanocratic gabbro to syenite types. Analcite is primary, deuteric, and hydrothermal. Also included is an intermediate trachytic and rhyolite group. Most of the varieties are soda-rich. Lonsdale shows the igneous rocks of the Terlingua–Solitario region to be closely related to those of the Spanish Peaks region and also to those of north-central Montana.

The analcite-bearing rocks obviously are a related series and originated through differentiation which preceded from melanocratic types through labradorite-rich types to syenite (Longsdale, 1940).

Baker (1935) has suggested that the uplifted block containing the Solitario dome is underlain by a batholith. In the adjacent sunken block in which nearly all the analcite-bearing rocks occur it is possible that the sinking resulted in rise of magma drawn from the lower and relatively basic part of the batholith. The result would be not a single immediate source of all the analcite-bearing rocks, but a number of differentiating masses in laccoliths and other minor intrusions from which, in the total, a relatively large number of varieties would be produced (Lonsdale, 1940). This is much the same arrangement as Larson postulates for the calc-alkalic series of the San Juan volcanic field.

The Chisos Mountains consists of a number of sharp peaks of intrusive and extrusive rocks. The area is referred to as an uplift, and is comparable to the Solitario in varieties of igneous rocks and includes alkalic types similar to the Terlingua–Solitario district.

Alkalic rocks have been penetrated in wells drilled for oil in the adjacent Delaware basin, but a problem exists in determining whether these are Tertiary or Precambrian (Flawn, 1952).

The west Texas alkalic province extends southeastward well into Mexico, for in the San Carlos Mountains an alkalic suite occurs. Kellum (1937) describes in the Sierra de San Jose division of the San Carlos Mountains an "alkalic rock complex," a feldspathoid-bearing sill, ijolite plugs, as well as micrognanite, quartz diorite, and diorite porphyries. There are also late basalt flows. The porphyries are probably laccoliths. In the Sierra de Cruillas division of the San Carlos Mountains Imlay (1937) describes microgranite and sills as the most common type of igneous rock. A vogesite sill was noted which is about 90 feet thick and at least 2½ miles long. A trachyte sill was also mapped. Basalt of alkalic varieties occurs as a laccolith and as sills and plugs. One plug is an

olivine basalt, the laccolith is an hauyne basanite, and some of the sills in one place are nepheline-hauyne basalt. The basalts were intruded considerably later than the micrograniles.

It is evident, in review, that the west Texas and northeastern Mexico alkalic province contains differentiates similar to the Spanish Peaks field of Colorado, the Rattlesnake Hills field of central Wyoming, and some of the igneous groups of central Montana. Fairly stable crustal conditions obtained in most all places, an olivine basalt was the parent magma, but probably some assimilation of alkalic country rock occurred, and in places a mixing of magmas in different states of differentiation seems to be necessary to explain the unusual types.

Calc-Alkalic Subprovinces

San Juan–Front Range Subprovince. The San Juan–Front Range will here include the igneous rocks of the San Juan Mountains, and the Front Range as well as the Spanish Peaks, Chico, and Raton basin fields (see map, Fig. 36.1). All the rocks of this large area have a notable calc-alkalic composition, range from basalt to rhyolite, and show a great variation from one flow to another.

San Juan Field. The great bulk of the San Juan Mountains volcanic field, about 100 miles in diameter, is made up of andesitic and rhyolitic rocks in about equal amounts. Basalts transitional to andesites are subordinate.

In the following stratigraphic sequence (Larson and Cross, 1956) the Miocene volcanics of the Potosi series are by far the most extensive and aggregate between 5000 and 6000 cubic miles in total bulk.

Quaternary andesite: one small body.
Erosion to mountain topography.
Pliocene (?) andesite, andesite-basalt, and rhyolite.
Erosion to peneplain.
Miocene latite-andesite.
Erosion to mountain topography.
Miocene (Potosi series) andesites, quartz latites, rhyolites, and subordinate andesitic basalts; several internal erosion intervals separating conformable sequences of lavas in which dominantly quartz-latite lavas and tuffs are succeeded upward by dominant andesites.
Erosion to mountain topography.
Upper Cretaceous to Eocene andesite (dominant), latite, and rhyolite; all occur locally and several internal erosion intervals can be recognized.

The volcanics lie partly on the northeast flank of a dome some 50 miles in diameter. They spread principally across the central part of the Uncompahgre Range of the Ancestral Rockies (Chapter 15). This range rose in Pennsylvanian time and was gradually buried during succeeding Permian, Triassic, and Jurassic time. In large parts of the range and the area upon which the volcanics accumulated not more than 3000 feet of strata existed, chiefly Cretaceous, prior to the Laramide doming. The area was characterized by doming on the west. To the east compressional deformation occurred in South Park and the Front Range (Chapter 25). At the time of Miocene volcanism large areas had been stripped of any sedimentary veneer, and the volcanics accumulated directly on the Precambrian crystalline rocks. The volcanics cannot be directly related therefore, to a basin of sedimentation, to a broad Laramide uplift, or to a belt of strong Laramide orogeny. As for the ancestral Uncompahgre uplift it would seem that its roots would long since have disappeared by isostatic adjustment before Tertiary volcanism occurred. This andesite assemblage is therefore somewhat of an anomaly but must not be neglected in shaping a theory of the origin of andesitic magmas in the orogenic belts.

Serial derivation from basic magma by fractional crystallization was the dominant process, but also prominent was the thorough mixing of magmas from the same common parentage but at different stages of differentiation. Some assimilation of country rock may also have occurred (Larsen and Cross, 1956).

> The evidence of mixing of magmas, contamination by foreign material, resorption of hornblende and biotite, and great variation in composition from one flow to another characterizes the San Juan volcanic pile . . . the evidence demonstrates that magmas of chemically related but quite dissimilar compositions, were generated locally within spongy subterranean chambers, and that

in general more than one chamber was tapped during an eruption (Waters, 1955).

Although the existence of an orogenic root is questionable, Waters suggests:

> . . . [This] part of the Rocky Mountain root was undergoing renewed granitization and anatexis, and that the volcanic rocks were fed from growing pods filled with mixtures of magma and migma. Pluto's genetic traits can actually be seen in the volcanic rocks! But it is not a root of argillites and graywackes that was undergoing partial melting as in the Cascades. Instead the richness in potash, and the abundance of biotite and hornblende in process of resorption point to a mountain root in a much later stage of metamorphic development—one in which the principal rocks were mica schists, amphibolites, and granodiorite intrusives.

Spanish Peaks Field. The igneous rocks of the Spanish Peaks area (Knopf, 1936) consist of two central stocks of which the older is a mass of granite porphyry. It is cut by the later pyroxene syenodiorite. A striking system of radial dikes (Chapter 25) evidently emanated from the stocks, and they range from highly silicic to mafic varieties. The order of intrusion is: 1, granite porphyry stock and granite porphyry dikes; 2, granodiorite porphyry stock and biotite porphyry dikes; 3, pyroxene syenodiorite stock; 4, microsyenodiorite; 5, teschenite, camptonite, and shonkinite, and trachydolerite; 6, augite syenodoirite porphyry; 7, camptonite and biotite lamprophyres. Their origin is discussed by Waters (1955) as follows:

> The order of intrusion in the stocks is the order of decreasing silica content, the reverse of the normal plutonic order. Noteworthy, too, is the great variety and abundance of the lamprophyres. Another interesting fact is that some of the lamprophyres are of the kind commonly considered related to calc-alkalic masses, whereas others are of the kind believed to be genetically related to alkalic rocks.
>
> Without added evidence from mineral paragenesis and inclusions it would be presumptuous to suggest that the Spanish Peak rocks may be igneous offshoots from a zone of biotite-rich metamorphic rocks that were undergoing partial fusion. Nevertheless, such a hypothesis, in contrast to derivation from a parental basalt magma, better fits the reversal in the "normal" sequence of intrusion. Also the rising temperature during anatexis, resulting ultimately in partial fusion of hornblende and biotite, can account for the formation of the varied suite of lamprophyres and can explain their heteromorphism (Waters, 1955).

Chico Field. In northeastern New Mexico adjacent to the Spanish Peaks field basalt flows cover over 700 square miles. They are here collectively called the Chico field. There were three periods of basalt extrusion separated by active stream erosion, and all are believed to be of Quaternary age although it is possible that the oldest is Pliocene. The basalt extrusions are mostly fissure-type eruptions, but some necks are noted. The extrusion loci have not been tied to post-Eocene structure. The volcanics occur on the southeast flank of the Raton basin of Cretaceaus and Tertiary age. Olivine basalts predominate in all three periods. The intermediate flows have the greatest variation and include olivine basalt, olivine-free basalt, olivine basalt with quartz inclusions, felspathoid basalts (tephrite?, olivine absent), and basanites (olivine present). Dacites, andesites, soda trachytes, and phonolites in minor amounts are also noted. All these rocks of the area probably originated from one magma whose original composition approximated olivine basalt. The suite is sodic alkalic (Collins, 1949; Stobbe, 1949).

The Chico field basalts are grouped in the same province with the San Juan andesites and rhyolites because they are adjacent and have had the same parentage, namely an olivine basalt magma. It should be noted that the Chico basalts are distinctly in a nonorogenic region.

Front Range Igneous Rocks. The transverse porphyry or mineral belt of Laramide intrusions and related rocks of the Front Range of Colorado has been reviewed in Chapter 25. The succession of igneous rocks and their chemical and mineralogical composition suggest that those of the western slopes were derived from an augite diorite magma, and those on the eastern slopes from an olivine basalt magma. The augite diorite magma differentiated into a series ranging from porphyritic diorite through porphyritic quartz monzonite to granite porphyry in comparatively shallow hearts. The olivine basalt magma gave rise to the differentiate series: diorite, monzonite, quartz monzonite, granite, alaskite, lead-silver ores; alkalic syenite, bostonite, pyrite gold ores; and biotite monzonite, biotite latite, latitic intrusion breccia, gold telluride ores, and tungsten ores. This was accomplished by withdrawal of portions of the changing residuum of the slowly solidifying magma into shallow reservoirs, and further differentiation by the subtractive processes of

crystal settling, crystal zoning, and filter pressing. The late differentiates of the olivine basalt magma were much more alkalic than those of the augite magma because of the initial difference in the composition of the parent magmas (Lovering and Goddard, 1950).

The eastern slope olivine basalt differentiate series is similar to that of the San Juan volcanic field except one is an intrusive succession and the other mostly an extrusive. Relative volumes are unknown, but at least, both are postulated to have come from an olivine basalt parent. Some assimilation may have occurred in the San Juan magma reservoirs but Lovering and Goddard, if the writer correctly understands, do not presume assimilation for the olivine basalt series of the east slopes of the Front Range However, the augite diorite parent magma may have been generated entirely by fusion of a crystalline basement rock.

Both magma subprovinces of the Front Range developed across the ancestral Colorado Range of Pennsylvanian age (see Chapter 25). This general area all through Paleozoic time had been dominantly positive and had received only a very thin veneer of sediments on the crystalline basement, which was broadly exposed by erosion as the Colorado Range was uplifted. The range was gradually buried during the Mesozoic, and Cretaceous beds were deposited on the Precambrian over wide areas of the old range and constituted in these places the only sedimentary rock at the time of Laramide orogeny. Again in Laramide times uplift was prominent but large-scale overthrusting occurred, especially in the western part of the old Colorado Range, now the Vasquez Mountains, the Williams Range, and the Gore Range, and the uplift, thrusting, and intrusive sequence are closely related in time.

Yellowstone Subprovince. The Yellowstone subprovince will here include the Absaroka, the Crazy Mountains, the Livingston and Adel Mountain fields as well as the Yellowstone Park field. The rocks of this province are generally calc-alkalic in mild contrast to the alkalic rocks of central Montana, previously described and also to the andesites and quartz latites of the Elkhorn Mountains volcanics and the Hogan formation. See Figs. 36.1 and 36.3. Actually the differences are slight and boundaries separating the three provinces are difficult to draw, principally because two of the volcanic centers have episodes of alkalic rock eruption separated by episodes of calc-alkalic rock eruptions. Superposed volcanic sequences are subprovinces in Larsen's nomenclature (1940).

Yellowstone Field. The eruptive rocks of Yellowstone Park range from basalt to rhyolite, with the basalts containing calcic plagioclase, augite, hypersthene, and olivine. The Absaroka Range has trachydolerites and orthoclase gabbros (alkalic) as its mafic rocks, and where the age relations have been determined the older effusives are generally calc-alkalic and the younger alkalic. The Absaroka field may therefore be placed in either the Yellowstone calc-alkalic province or the central Montana alkalic province.

Crazy Mountains Field. The igneous rocks of the Crazy Mountains consist of an older calc-alkalic series of two stocks and associated dikes, sills, and laccoliths, and a younger alkalic series, found chiefly in the northern part of the mountains, and occurring as sills, laccoliths, and dikes. The alkalic bodies are richer in soda than any of the other groups of central Montana, and have been determined as granite porphyry, syenite, nepheline syenite, shonkinite, and lamprophyre. The older and more calcic stocks are chiefly diorite with minor amounts of granodiorite, gabbro, and picrite (Larsen, 1940).

Livingston Field. The Livingston formation is a series of pyroclastic rocks several thousand feet thick which crop out on the north side of the Beartooth Mountains. They grade laterally into the Claggett, Judith River, Bearpaw, and Lennep formations and hence represent a center of volcanism that was active during most of the Montana epoch of the Upper Cretaceous. Pyroxene andesite breccias are by far the most abundant, and occur both above and below hornblende andesite breccias (Vhay, 1939).

Adel Mountain. A fairly large volcanic field in the southern end of the Foothill belt of the Canadian and Montana Rockies, west of the Highwood Mountains and on the northern end of the Big Belt Mountains may be divided into two parts. Its southeastern part, the Adel Mountain, has been studied by Lyons (1944). He finds that the volcanic rocks consist of potash-rich basalts which were erupted on Cretaceous sediments. The trachybasalt breccias, flows, and agglomerates are 3200 feet thick and have been intruded by many chonoliths, sills, and dikes ranging from gabbro to quartz monzonites. The chemical and mineralogical analyses relate

the field to the Highwood Mountains and the alkalic province of central Montana.

Summary. The Yellowstone subprovince as a whole is one of olivine basalt parentage and although it is in a Laramide belt of mild to appreciable deformation, it is definitely not of the orogenic andesite lineage; it belongs to the nonorogenic calc-alkalic and alkalic provinces east of the central Rockies.

Crowsnest Volcanic Field. Still another Cretaceous volcanic field has been described in the deformed belt (MacKenzie, 1956). It is known as the Crowsnest volcanics and occurs about 30 miles north of the international boundary (see Figs. 20.2 and 20.6). The volcanic deposit lies within the mid-Cretaceous sediments; viz., the continental Blairmore (Dakota) formation underlies it, and the marine Blackstone (Benton) overlies it. At Coleman the lower unit consists of trachyte agglomerate beds. This is overlain by ash beds with scattered large fragments of pyroclastics; this in turn is overlain by water-laid ash beds rich in andesite, and this in turn by more ash beds with varying amounts of coarse pyroclastics. Some of the ash beds are hard resembling flows; but no actual flow rocks are reported. Four main lithologies have been identified, namely, augite trachyte breccia, tinguaite, andesite tuff, and phonolite tuff. The name blairmorite has been suggested for certain analycite-rich fine-grained rocks in the volcanics.

> The rock on account of its ultra-alkaline nature, will show numerous variations in texture and in proportions of component minerals . . . (MacKenzie, 1956).

It is evident that the Crowsnest volcanics are alkalic and related to the central Montana petrographic province rather than a field to the south in the Foothill belt of Montana, the Dearborn River which is generally andesitic. The Dearborn River field is described under a later heading.

The age by stratigraphic position is Mid-Cretaceous, and by potassium-argon dating (Folinsbee *et al.*, 1957) is 96 m.y. The Crowsnest volcanics are slightly older than the Dearborn River Volcanics, according to stratigraphic position, and slightly younger than the main Nevadan batholithic intrusions to the west.

The Crowsnest volcanic field is estimated to have a maximum average thickness of about 1000 feet and to spread over 500–600 square miles. It contains about 50 cubic miles of rock.

Southern Colorado Plateau Basalt–Rhyolite Province

Extent. Several volcanic fields along the southern margin of the Colorado Plateau may be conveniently grouped together because of their close proximity, but they hardly have enough common characteristics to justify the grouping. The questionable province starts on the east with the Jemez field in north-central New Mexico and includes the Mount Taylor, the Datil, the San Francisco, and the Uinkaret fields. See map, Fig. 36.1.

Characteristics. The Jemez, Mt. Taylor, and San Francisco fields are the result of large, central vent-type volcanoes. One large volcano or a cluster of several with numerous, later, small cinder and lava cones make up the fields. The rocks range from basalt to rhyolites and appear to represent the basalt-rhyolite differentiation suit. The source of the lavas in the large Datil field is not recorded in the literature as far as the writer can determine. The Uinkaret field consists of youthful small cones and basalt flows.

Description of Fields. The Mt. Taylor field is dominated by the Mount Taylor volcano which erupted in late Miocene time, after folding and faulting in the district.

> The volcano broke out in a syncline. The eruption, which occurred in a fairly well defined compositional sequence, began with rhyolitic tuff. This was followed by relatively quiet eruptions of porphyritic lavas in which two and possibly three series are distinguishable on the basis of their content of potash feldspar. The oldest of these is porphyritic trachyte, but the volume is very small. The next eruption was a large volume of porphyritic latite, interrupted, however, by at least one more flow of porphyritic trachyte. The latite, in turn, was followed by a slightly smaller volume of porphyritic andesite.
>
> The total volume of the tuffs and lavas is about 12.5 cubic miles, of which about 5 cubic miles is rhyolitic tuff, 4 cubic miles is latite, and 3.6 cubic miles is porphyritic andesite.
>
> The erosion surfaces that subsequently were developed around the base of the cone later became flooded with sheets of nonporphyritic basaltic and andesitic lavas erupted from the scores of vents that comprise the volcanic field. A few of the sheets were erupted prior to the latest eruptions on Mount

Taylor, but most of them were erupted after Mount Taylor had become quiescent and they overlap the outer edges of the Mount Taylor cone (Hunt, 1938, p. 58).

The Mount Taylor central vent volcanics are slightly more alkalic than the rocks of the laccolithic groups of the Colorado Plateau, and Hunt, therefore, points out a close tie of the two.

The San Francisco volcanic field is much larger than the Mt. Taylor, and the initial activity consisted of the eruption of about 30 cubic miles of sheet basalt over a broad structural dome, the Coconino Plateau. Several large vent-type volcanoes broke out; San Francisco Mountain being built of almost 40 cubic miles of volcanic ejecta, Kendrick Peak of more than 6 cubic miles, Bill Williams Mountain of 3 cubic miles, and O'Leary Peak of 2 cubic miles. The five stages of eruption of San Francisco Mountain volcano were as follows: 21 cubic miles of latitic lava, tuff, and breccia, 13 cubic miles of pyroxene dacite lava, 0.5 cubic mile of hornblende dacite, 0.5 cubic mile of rhyolite, and 3 cubic miles of andesite. On a succession of erosion surfaces four separate basalt flows occurred, and basalt lavas and pyroclastics were extruded from about 200 small vents. This was the last phase of activity and about 20 cubic miles of lava was extruded. One of the cinder cones, Sunset Peak, was active 800 years ago (Robinson, 1913).

Datil field of eastern Arizona and adjacent New Mexico, is largely andesite, with subordinate amounts of rhyolite, rhyolite tuff, quartz latite, and various pyroclastics consisting mostly of basalt (Sabins, 1957). The Mogollon Mining District is within this large field, and there Ferguson (1927) describes 8000 feet of andesite, rhyolite, rhyolite tuff, and quartz latite. This assemblage savors of the Great Basin latite-monzonite province, and perhaps has some welded tuffs. Variations from mostly basalt to mostly latite and rhyolite would appear to be dependent upon the amount of silicic crust effected by partial melting. Refer to theory presented under next heading, Basin and Range latite-monzonite province.

Basin and Range Latite-Monzonite Province

Extent and General Characteristics. Butler (1920) summarized the volcanic rocks of western Utah and adjacent parts of the Basin and Range physiographic province as follows. They range in composition from rhyolite to basalt, but the great bulk of the material is of intermediate composition, including rather basic rhyolites, quartz latites, dacites, and andesites. Basalt is very subordinate in amount when compared with the series as a whole though present in many localities and usually conspicuous as representing the latest volcanic outflows.

A large region in Nevada and western Utah consists dominantly of welded tuffs of approximate quartz-latite composition. The alkalic types of rock, such as the leucite and nepheline-bearing lavas are to the writer's knowledge, very scarce, and have only been noted in East Fork Canyon of the High Plateaus where Dutton described an isolated occurrence of phonolite and in the Traverse volcanics of the Oquirrh Mountains (Gilluly, 1932). A brief review of the Tertiary volcanic rocks in southern Arizona indicates that they are essentially the same as in Nevada and western Utah, and fit Butler's general description. The intrusive rocks are principally in the form of stocks. They have a dioritic to granitic composition, with monzonitic the most common. Like the extrusives the intrusives are preponderantly intermediate to acidic in composition.

Intrusive Rocks. The following is a tabulation of the intrusive rocks of ninety-five mineral districts which Stringham has made in the course of a systematic study of the mineralized and barren stocks of western Utah, Nevada, Arizona, California, and New Mexico (personal communication). About 300 plutons, mostly stocks, are shown on various maps of areas in these states or are known from personal field work, according to Dr. Stringham. When a district is mapped, more intrusive bodies are usually found, so he estimates that possibly 1000 intrusions may exist. In western Nevada most of these are probably satellites of the Sierra Nevada batholith and not Laramide or later in age as elsewhere in the Great Basin.

The intrusive bodies shown in the table on page 574 are charted on Fig. 36.3, where it is seen that the three classifications of the tabulation have little significance geographically. It might be concluded that western Utah and eastern Nevada are free of intrusions as basic as diorite, but elsewhere in the Great Basin the three divisions are fairly well scattered.

The Tertiary intrusives of central New Mexico range from diorite to granite, with a preponderance of monzonite and quartz monzonite (Lind-

Intrusive Bodies of Mineral Districts in Great Basin

Diorite and Andesite	Granodiorite Quartz Monzonite, Monzonite, Latite, Dacite	Granite, Rhyolite
	Arizona	
Pearce (1)	San Manuel (3)	Morenci (18)
Superior (2)	Christmas (4)	Bisbee (19)
	Big Bug (5)	Oatman (20)
	Johnson (6)	Kofa (21)
	Miami (7)	Mammoth (22)
	Ajo (8)	Jerome (23)
	Tombstone (9)	Ray (24)
	Silver Bell (10)	Chloride (25)
	Castle Dome (11)	Arivaipa (26)
	Courtland-Gleeson (12)	Congress (27)
	Patagonia (13)	Aqua Fria (28)
	Helvitia (14)	
	Harshaw (15)	
	Bagdad (16)	
	Twin Buttes (17)	
	New Mexico	
Mogollon (29)	Kingston (32)	Lordsburg (38)
Steeple Rock (30)	Central (33)	
Pinos Altos (31)	Magdalena (34)	
	Tyrone (35)	
	Hillsboro (36)	
	Chloride Flat (37)	
	Nevada	
Tonopah (39)	Tybo (49)	Pioche (62)
Aurora (40)	Copper Canyon (50)	Goodsprings (63)
Tuscarora (41)	Candelaria (51)	Jarbidge (64)
Fairview (42)	Gold Acres (52)	Bristol (65)
Divide (43)	Ely (53)	Silver Peak (66)
Virginia City (44)	Yerington (54)	Wonder (67)
Searchlight (45)	Goldfield (55)	Rochester (68)
Eldorado (46)	Eureka (56)	Unionville (69)
Tenabo (47)	Copper Basin (57)	National (70)
Lewis (48)	Austin (58)	Seven Troughs (71)
	Mill City (59)	Bullfrog (72)
	Getchell (60)	Delamar (73)
	Cherry Creek (61)	Hamilton (74)
	Nevada	
		Manhattan (75)
		Round Mtn. (76)
		Bullion (77)
	California	
Bodie (78)	Cerro Gordo (79)	Calico (82)
	Randsburg (80)	Blind Spg. Hill (83)
	Mohave (81)	Ludlow (84)
		Darwin (85)
	Utah	
Park City (86)	Stockton (87)	Ophir (91)
	San Francisco (88)	Mercur (92)
	Gold Hill (89)	Bingham (93)
	Lucin (90)	Marysvale (94)
		Tintic (95)

gren *et al.*, 1910). They are intruded into Precambrian granites and schists and all parts of the Paleozoic and Mesozoic stratigraphic sequence, and take the form of stocks, sills, and dikes. The volcanics are basalts, andesites and rhyolites, with the order of eruption generally, rhyolite, andesite (or latite), rhyolite again, and finally basalt.

Possibly the largest Laramide or Tertiary intrusive in Utah, Nevada, Arizona, or New Mexico is that of the Sierra Blanco in Lincoln County, New Mexico. It is probably connected underground with plutons to the north in the Jicarilla Mountains and Gallinas Mountain and to the east in the Capitan Mountains. The Sierra Blanco, Jicarilla, and Gallinas plutons extend a distance north-south of 70 miles, and the Sierra Blanco pluton itself has a maximum width of 15 miles. The Capitan pluton extends 20 miles in an east-west direction and about 5 miles in a north-south direction. The major intrusive mass, the Sierra Blanco, and the Jicarilla, have penetrated a basin downwarp containing Cretaceous strata not much larger than the exposed plutons. The Gallinas and Capitan plutons have

come up through Pennsylvanian and Permian strata which are nearly horizontal.

Another fairly large pluton is one in the Organ Mountains, about 30 miles north of El Paso. It extends 18 miles north-south and 9 miles east-west.

Lindgren *et al.* (1910) describe the large plutons mentioned above as monzonite and quartz monzonite porphyries. They observe that the intrusive monzonites and effusive latites and andesites in general in the central north-south belt of New Mexico have a fairly uniform composition and suggest that all were derived from an intermediate magma. The last differentiates were basalt and rhyolite, which were the last ejections.

The general problem of the nature of the primary magmas will be discussed presently, but it should be said that the fusion of a gneissic, schistose, and granitic basement such as would produce a monzonitic magma, would not have enough magnesium and iron in it to yield a basaltic differentiate, especially an olivine basalt. Also the volume of basalt in some fields is too much to have been derived from the postulated parent primary monzonitic magma.

Extrusive Rocks. As for the extrusive rocks only limited information is at hand. Those of central and eastern Nevada and southwestern Utah consist of a thick older assemblage and a thinner younger group which is approximately equal in age to the younger volcanics of western Nevada, southern California, Arizona, and southwestern New Mexico. The older assemblage is dominantly of the quartz latite type, and more conspicuously, it consists chiefly of a great series of avalanche or welded tuff deposits, whereas those in peripheral location are more of the basalt-andesite-dacite-rhyolite suite.

Brief descriptions of selected fields outside of the avalanche subprovince follow.

In the Ajo District of south-central Arizona the lavas are basaltic, andesitic, and latitic. In southern Nevada in the Goodspring's quadrangle the extrusive rocks range from latite through andesite to basalt. At Goldfield, Nevada, the eruptive sequence is rhyolite, latite, rhyolite, olivine basalt, andesite, dacite, andesite, rhyolite, andesite, olivine basalt, rhyolite, and olivine basalt. At Gold Hill on the Utah–Nevada line a normal series of basalts, andesites, and rhyolites occurs. They are all rich in potash.

Latite Welded Tuffs Subprovince. The welded tuff subprovince is shown on the maps of Fig. 36.1 and 36.3, and its existence has only recently become clear. Mackin and Cook and associates in southwestern Utah and several petroleum and U.S. Geological Survey geologists have recognized the welded tuffs (ignimbrites) and something of their magnitude. However, the writer is especially indebted to Dr. Howel Williams for the following résumé. He was among the first to gain the conception of the unique field and the almost unbelievable magnitude and awesomeness of the eruptions.

Welded tuffs are formed during eruption by distention of magma in which the vapor tension is low. Instead of explosive eruption of vitric ashes, the discharge is in the form of a glowing avalanche that sweeps rapidly downslope. The most widespread of these result from escape of foaming magma through swarms of fissures as a mixture of incandescent spray, droplets, and larger clots enveloped in hot, expanding gas. So mobile are these mixtures that they spread over vast areas, down even the gentlest gradients. Other glowing avalanches issue from the flanks of volcanic domes of Pelean type; still others originate when foaming magma is upheaved en masse until it spills over a crater rim and then, aided by gravity, races downward. Because these avalanche deposits accumulate rapidly and usually to great thickness, many remain hot for a long time, especially in their central parts. As a result, the shards of glass, while still hot and under heavy overburden, are squeezed and flattened, and some are buckled between phenocrysts. At the same time pumiceous lapilli and bombs are deformed to disks, some of them paper-thin, and all the constituents become firmly annealed. The rocks thus formed are called welded tuffs. They have a delicate, streaky lamination deceptively like the fluidal banding seen in many lavas. Besides, some of them develop columnar and spherulitic structures as they cool, so that their resemblance to lavas is increased. Little wonder, therefore, that welded tuffs have often been wrongly identified. The fact is that they are now known to be of truly colossal extent in the circum-Pacific volcanic regions, and they are undoubtedly widespread elsewhere (Williams *et al.*, 1954).

The welded tuffs according to Williams constitute 95 percent of the older and more voluminous volcanics of the avalanche subprovince. They probably average over 2000 feet thick, and south of Eureka, Nevada, they are 8000 feet thick. This general area is the part of the field of maximum thickness. Since immense amounts of these easily weathered tuffs have been removed, the original volume was undoubtedly larger than

present thicknesses indicate. At least 30,000 cubic miles of welded tuffs were erupted in this subprovince.

They originated in fissure eruptions, and dike feeders are the rule. No cones, central vents, radial dike swarms, or quaquaversal dips have been noted. Dr. Williams' ideas about the age of the welded tuff accumulations is as follows. The rapidity of accumulation is startling because a layer as thick as 1000 feet may have accumulated in one day. Although there were many eruptive centers the entire explosive activity occurred undoubtedly in a very short time geologically, say a few thousand years. The major activity of a certain group of fissures may have taken place in three or four days. This is deduced because of the absence of erosional breaks of any kind in the succession of welded tuff flows. Soil profiles between flows were sought but not found.

A potassium-argon age determination yielded a date of 35 ± 2 m.y., and Professor Williams thinks this will prove to be the age of the main unit of the thick assemblage of welded tuffs over the entire subprovince. There are younger welded tuffs above and outside the subprovince, but these are another matter. The age would then be early or mid-Oligocene according to which absolute time scale is used.

According to E. F. Cook (personal communication), who has studied the welded tuffs extensively through Nevada and western Utah and who has attempted to correlate many measured sections, breaks in the depositional sequences occur, with flows and sediments interlayered. He believes the eruptions were intermittent and extended through the Oligocene into the Miocene, and suggests that the period through which the welded tuffs were erupted was several million years long. T. B. Nolan informed the writer that the 35 (or 34) m.y. potassium-argon date appears to conflict with Miocene fossils in the Eureka–Austin–Winnemucca area. Since the interest of a number of geologists and geochemists is high on the problem, our knowledge will undoubtedly be more precise in a short while.

The surface at the time of the numerous and widespread outbreaks seems to have been very flat, according to Dr. Williams, because wherever the base of the series is exposed it is without relief, and since the individual avalanche flows can be traced scores of miles, there must not have been sizable topographic obstructions in their way. This is especially true for the upper flows of the welded tuff sequence, according to Dr. Cook, but he believes the early flows filled basins of appreciable relief or closure. The bulk of the material of the flows is approximately of quartz latite composition, and it is mostly slightly potassic with some parts rather potassic. Some inclusions occur and these confirm the suspicion of Professor Williams that assimilation of the crystalline basement of the silicic crust is involved in the origin of the welded tuffs.

Eleven ignimbrites are widespread in southeastern Utah and have been given formal stratigraphic names by Mackin (1960). He says:

> The fact that the oldest of them lies unconformably across the beveled edges of thrusts and folds involving late Cretaceous strata indicates that the beginning of volcanic activity post-dates the Laramide orogeny. As planar units which provide a record of Tertiary crustal movements, the ignimbrites confirm the Gilbert theory, based originally on physiographic evidence, that block faulting has been the characteristic type of post-Laramide deformation in the Great Basin.

The volcanic field of southwestern Utah abounds in welded tuffs. These and associated volcanics are described by Cook (1957) in probably the best account of them so far. Indurated acidic pyroclastic rocks ranging from welded tuffs to bedded tuff-breccias dominate the volcanic column which is several thousand feet thick. The bedded tuff-breccia occurs in beds 2 to 20 feet thick and fills depressions in a rough topography developed on folded and faulted volcanic rocks. Its tuffs are nonwelded but in one locality appear to grade downward into welded tuff-breccia. Cook concludes that the bedded tuff breccia was deposited by a series of rapidly moving, widely spreading flows of gas (possibly steam), hot water, and solid particles, in which the temperature was below that required for welding.

Breccia and air-fall tuff form a minor portion of the volcanic rocks of the area. Lava flows, conspicuous locally but also of minor amount, include dacite porphyry, locally porous and fluidal, latite (?) and latite porphyry, olive-brown to black andesite, and dark gray to black basalt. The higher part of the Pine Valley Mountains consist of latite (or quartz latite) porphyry. Except for a chilled basal phase the latite is lithologically uniform throughout its thickness of 2000 feet, and is similar mineralogi-

cally and texturally to the monzonite porphyry of the intrusive laccolith there.

The mode of origin of the extrusive rocks of the Pine Valley Mountain appears to be related to their chemical composition. The basalts, andesites, and latites are flows; the dacites are found both as ignimbrites and as flows; and the rhyolitic rocks are all ignimbrites. Apparently the more acidic magma effervesced into *nuees ardentes,* although some of dacitic composition merely foamed into frothy flows; the intermediate and basic lavas, on the other hand, welled up without violent loss of gas to form finely vesicular flows. (Cook, 1957, p. 49).

A study by Van Houten (1956) of the Cenozoic sedimentary and related volcanic rocks of Nevada and western Utah indicates that a good datum for correlation is a tuffaceous unit of late Miocene and early and mid-Pliocene age. A vitric tuff in this general bentonitic and tuffaceous unit is prominent and widespread. It rests on somewhat older Oligocene and Miocene (?) volcanic rocks in southern, central, and western Nevada, as well as locally in the northeastern part of the state.

The lower volcanic units were tilted by fault block rotation and eroded before the widespread, late Miocene–Pliocene tuffaceous unit began to accumulate. During this late Miocene to mid-Pliocene interval the southern Cascade andesites were accumulating as well as the younger basic lavas of the Sierra Nevada. The inference is natural that the volcanism and faulting are related, but this subject will be left until later for discussion.

The study by Van Houten emphasizes the existence of extensive fluviatile and lacustrine deposits derived largely from the eruptive centers. The sedimentary derivatives fill the numerous intermontane valleys in places to the depth of over 5000 feet and have been tilted in the Basin and Range faulting to be exposed on the backs of the tilted blocks. In some places as much as 10,000 feet of volcanic rock, including derived stratified outwash and lacustrine deposits have been measured (personal communication, various petroleum geologists). The volcanic fields, other than the avalanche subprovince, have not been determined and circumscribed. Those shown on the map of Fig. 33.7 are taken from the *Geologic Map of the United States* (1932), and it is presumed they are the most continuous and thus indicate the major centers of volcanism. It is evident that this representation is likely to be changed considerably by future work.

As a result of the extensive exploratory work for oil in Nevada, nearly every intermontane valley is regarded as a downfaulted block, but only in a few places have the faults been shown on maps either published or available to the writer. The best recourse, it seems, is to show each valley by a single fault, and this has been done in the absence of better information. The Basin and Range fault system is undoubtedly more complicated than shown.

Origin of the Latite Magmas. In 1932, Gilluly observed from a study that focused in the Oquirrh Mountains that a close relation of all the extrusions is evident, and although several of the volcanic masses have been described as andesites, these when analyzed have the alkali ratios characteristic of latites. Whether they contain orthoclase or not, their chemistry justifies the inference that they all belong to the latite and quartz latite group.

> Plotting the CaO, K_2O, and Na_2O as ordinates and the silica as abscissas reveals the interesting fact that the soda shows almost linear decrease with increase of silica. The average slope of the curve of soda is almost precisely that which would appear from the mere addition of silica to the monzonite magma. However, the lime decreases at a rate altogether disproportionate to the silica content, and the potash remains very nearly constant or decreases at a much lower rate than the silica increases. This relation is close to that which would be expected as a result of differentiation of a monzonitic magma by fractional crystallization.
>
> Both intrusive and effusive rocks of western Utah have CaO, Na_2O, and K_2O proportions close to the average quartz monzonite, and hence are believed to be of the monzonite kindred. Wherever chemical analyses are available, it is seen that the so-called andesites are without exception so high in potassa that they are properly classed as latites. Similarly, several so-called dacites resemble quartz latite closely (Gilluly, 1932, pp. 66, 67).

Gilluly concluded and reaffirms the conclusion in recent correspondence that there is no evidence here of any more basic rocks that could be considered parental to the latite-monzonite magmas.

> In Bowen's scheme of magma evolution it would be necessary to have basaltic and andesitic rocks prior to the quartz monzonite. All of the volcanics

and plutonics, with the trivial exception of some alkaline bodies, are either quartz latite or quartz monzonite. In other words there is no evidence of any magma antecedent to these. In that sense the monzonite is primary magmatic rock.

I know of no discussions of this problem in the literature. I would be inclined to feel that the magma is perhaps a regenerated one, produced by melting of the deeper part of the crust (James Gilluly, personal communication, 1957).

In a detailed study of the Bingham stock Stringham (1953) finds types other than monzonite according to mineral content, such as granite, monzonite, diorite, syenite, and syenodiorite, with granite the dominant variety. Chemically, however, the granite is not much different from quartz latite or quartz monzonite. The granite and actinolite syenite appear to have originated by granitization, perhaps as fringe effects of a central or more deep-seated magma, according to Stringham.

Western Montana and Eastern Idaho Andesite-Granodiorite Province

Peculiar Nature of Province. The term, andesite-granodiorite province, is used for want of a better one for the assembly of igneous rocks in southwestern Montana and adjacent Idaho. The province does not fall readily into any of the other categories outlined at the beginning of this chapter. It seems to be a hybrid of the Nevadan batholithic province and the monzonite-latite province of the Great Basin with certain aspects of the andesite province added.

Several volcanic fields of Late Cretaceous age are known in the general province of western Montana as shown on Fig. 36.1. They may be grouped into the Livingston, the Elkhorn Mountains, and Adel Mountain fields and the Hogan formation. The Livingston and Adel Mountain fields are classed under the Yellowstone subprovince of calc-alkalic rocks, whereas the Elkhorn Mountains and Hogan fields are andesites and latites, and therefore less calc-alkalic. They belong more properly to the orogenic belt.

Elkhorn Mountains Field.

Remnants of a thick plateaulike accumulation of calc-alkaline volcanic rocks of probable Late Cretaceous age—the Elkhorn Mountains volcanic rocks—are exposed in an area of about 3700 square miles around the Boulder batholith in the Elkhorn Mountains and Boulder Mountains, western Montana. The presence of similar rocks across the Jefferson River to the south, and near Wolf Creek to the north, suggests that the volcanic pile once covered more than 10,000 square miles.

In places these rocks rest unconformably on Paleozoic and perhaps older rocks. Elsewhere they are gradational into underlying tuffaceous sedimentary rocks of Late Cretaceous age, and the contact is arbitrarily placed at the base of the lowest volcanic conglomerate, breccia, or flow.

The volcanic pile comprises three major units; maximum thickness of each exceeds 5000 feet. The lower unit consists predominantly of dacitic, andesitic, and basaltic fragmental rocks and autobrecciated lava flows. The middle unit is about half quartz latite in welded tuff sheets as much as 300 feet thick, interlayered with more calcic bedded pyroclastic rocks and autobrecciated lava flows; it is locally unconformable on the lower unit. The upper unit consists dominantly of reworked volcanic rocks and subordinately of fine-grained pyroclastic rocks. A thick succession of basalt flows near Elliston, Montana, may be equivalent to the upper part of this unit or may be younger.

The volcanic rocks were altered by and locally foundered in penecontemporaneous shallow-magma reservoirs. They were folded and faulted and later invaded and thermally metamorphosed by the Boulder batholith (Klepper and Smedes, 1959).

Regarding the intrusive rocks, Klepper *et al.* (1957) say:

The intrusive igneous rocks, except for a few felsite dikes of uncertain age, are divisible into two groups, primarily on the basis of structural relations and secondarily on the basis of composition and fabric. The older group of dioritic and andesitic rocks were intruded in part, if not wholly, prior to the main folding and are similar in chemical and mineralogical composition to the Elkhorn Mountains volcanics. They were probably emplaced throughout the period of volcanism that commenced in late Niobrara time and continued until late Cretaceous time. The younger group consists chiefly of quartz-bearing phanerites but includes rocks ranging from gabbro to alaskitic granite and aplite. These rocks were emplaced after the main episode of folding and faulting. The Boulder batholith, composed dominantly of quartz monzonite, is the principal body of this younger group (Klepper *et al.*, 1957).

Although Klepper and Smedes class the Elkhorn volcanics as calc-alkalic, the rocks are less alkaline than the Livingston and Adel Mountain volcanics and, with the associated intrusive rocks, are more closely related to the igneous rocks of the Great Basin than to those of central Montana.

Hogan Field. The volcanic rocks of the Hogan formation, according to George Viele (Ph.D. thesis, University of Utah, 1960), are nearly 2500 feet thick and consist of interbedded breccia, welded tuff, volcanic-rich graywackes, shale, black sandstone, and arkose. Andesitic and more acidic

eruptives provided the material for the pyroclastics. The rocks are not rich in the alkalies and hence stand apart from the nearby Adel volcanics.

The nearby Adel volcanics, according to Viele, are probably of St. Mary River formation age (latest Cretaceous) and extend into the Paleocene. The Hogan volcanics are slightly older and are correlated with the upper part of the Two Medicine formation and the lower Horsethief formation. The Adel field has been only slightly tilted; whereas the Hogan field beds have been involved in the folding and thrusting of the Foothill belt.

Batholiths and Stocks. The region is noted for its large intrusives, particularly the largest, the Boulder batholith. This cluster of intrusives in west-central Montana and adjacent Idaho is easily the most voluminous anywhere in the Laramide belts. The plutons consist dominantly of quartz monzonites and granodiorites, although a number of variations in facies and separate intrusions have been noted which range from gabbro to granite. It has been pointed out by Emmons and Calkins (1912) that the intrusives of the west-lying Philipsburg district are less alkalic than the Boulder batholith and that the Boulder batholith has mild alkalic affinities to the high alkalic rocks of the central Montana petrographic province. This may mean that fusion of the crystalline basement is to be reckoned with, and that the fused rocks become less alkalic westward.

The Boulder batholith has been studied by Profesor Knopf (1957). He describes it as follows:

> On the basis of a recent potassium-argon age determination, which gave 87 million years as the most probable age of the granodiorite of the Boulder bathylith, it is concluded that the bathylith was emplaced late in Cretaceous time.
>
> The Boulder bathylith has hitherto been considered to be a one-magma intrusion, but like other large plutonic masses it proves to be of composite construction. The order of intrusion is (1) Unionville granodiorite, a basic hypersthene-bearing granodiorite which itself has developed basic faces of granogabbro; (2) Clancy granodiorite; (3) porphyritic granodiorite; (4) biotite adamelite; and (5) muscovitic biotite granite. Alaskite and aplite are abundant and were presumably (but not yet proved) developed most abundantly during the final stages of bathylithic consolidation. The order of emplacement of the successive intrusives is in the order of increasing silicity.
>
> The Boulder bathylith and its satellitic stocks, have exerted extensive contact metamorphism, both purely thermal and pyrometasomatic. Most notably, the Helena dolomite has been transformed into aphanitic tremolite-diopside hornfels to a maximum distance of 10,000 feet from the edge of the bathylith. The highest rank of metamorphism attained is in sillimanite-cordierite-microperthite hornfels, remarkable rocks that have formed at widely separated localities. In places the magma has reacted with limestone xenoliths with the result that the xenolith is surrounded by an aureole of augite granodiorite. In other places the evidence appears to demand that the magma in depth had dissolved limestone. By this syntexis alkalic rocks were generated that range from mildly alkalic, such as the Priests Pass leucomonzonite and the syenodiorite of the large stock northwest of Helena, to strongly alkalic, as represented by the nepheline shonkinite occurring east of Montana City.

In order to be consistent with the epoch designations for the absolute ages of the Sierra Nevada batholiths, we must assign a mid- or early Late Cretaceous age to the 87 m.y. date by the potassium-argon method of the Boulder batholith. Since the batholith is composite we wonder whether an early or late pluton in the intrusion cycle there is dated. It should be noted also that a date of about 103 m.y. by the lead-alpha activity ratio method is assigned to the Idaho batholith, but again, it is not known what part of the batholithic cycle is dated. See Chapter 21. The difference in method used also leaves the comparison uncertain. For the time being, however, we should presume that the Boulder batholith and associates in western Montana are slightly younger than the Idaho batholith.

Conclusions. The batholiths and stocks are fairly similar in composition to those of the Great Basin but have overtones of similarity in their variations and size with the major batholiths of the Nevadan belt. The volcanics are more andesitic than the dominantly latitic volcanics of the Great Basin. It seems, therefore, that the western Montana and eastern Idaho province displays characteristics transitional from the moderately orogenic region of the Basin and Range province to the intensively orogenic region of the Pacific marginal regions.

Late Precambrian (?) Sills and Flows. Sills, flows, and some dikes of basic rock have been noted in a number of places in the region of the Boulder batholith and northward through the Garnet Range to Glacier National Park. The sills are all intrusive into the upper part of the Belt series and appear as dark beds of remarkably uniform thickness for many miles. They also hold remarkably well to a single stratigraphic horizon and range up to 300 feet thick.

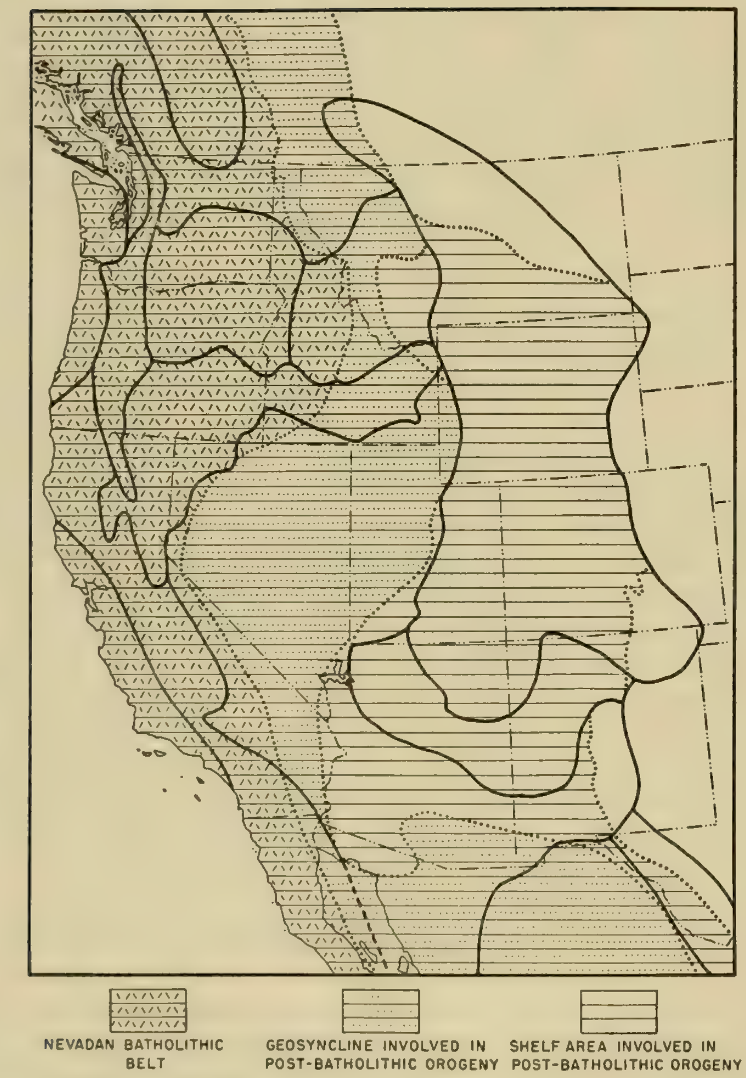


Fig. 36.6. Relations of tectonic provinces of western United States to igneous. To identify igneous provinces compare with Fig. 36.3.

The sill in the upper Belt series of Glacier National Park is a dark diorite but in places is pink due to orthoclase feldspar and in other places green from the presence of mica. Stratigraphically several hundred feet above the sill is the Purcell lava unit 50 to 275 feet thick. It is made up of several submarine flows with pillow structure. Hundreds of feet higher forming the top of the exposed Belt strata are several other flows. It is not known whether the flow rocks are spilites and keratophyres (Dyson, 1949).

Basic sills of diabase (?) occur in the Ray–Superior–Globe area of the Mountain Region of south-central Arizona, and are the only other ones known to the writer in the Laramide systems similar to those in the Belt series of western Montana.

RELATION OF TECTONIC TO IGNEOUS PROVINCES

The major tectonic and igneous provinces of the western United States are related to each other on the map of Fig. 36.6. The Nevadan batholithic belt has about the same distribution as the previously existing eugeosyncline. The Mexican geosyncline has considerable volcanic material on the north and west in the Cretaceous sediments, but the Nevadan belt of metamorphism and batholithic intrusions developed to the west in Baja California. The later second cycle batholiths intruded the western flank of the geosyncline, however, in great volume.

The batholithic belt was the site of later basaltic and andesitic eruptions in two regions: (1) a narrow zone extending from northern California to southern British Columbia, and (2) in central British Columbia along the east-central part of the broad batholithic belt with the vast Coast Range batholith entirely on the west. As in the Sierra Madre Occidental no row of late Cenozoic stratovolcanoes occurs in the broad fields of central British Columbia. The Cascade basalt-andesite field with its row of stratovolcanoes correlates in north-south extent with the eastern bulge of the batholithic belt in Oregon, Washington, Idaho, and southern British Columbia. It does not continue southward where the batholithic belt narrows in central and southern California. A genetic relation to the bulge is implied.

The igneous rocks east of the batholithic belt in the miogeosyncline of Nevada and western Utah are mostly of the monzonite-latite clan with

numerous stocks and widespread volcanism. The magma has generally risen through a thick sedimentary veneer, and little basalt has emerged at the surface. However, similar intrusions and extrusions occur in southern Arizona where the sedimentary rocks are thin, so the sedimentary veneer is not important, it seems. The latite magma is considered to be a primary one, and its origin will be taken up on later pages.

The Laramide Rockies of central Montana, Wyoming, and Colorado, as well as the Colorado Plateau constitute a large calc-alkalic and alkalic province where assimilation of calcium-, sodium-, and potassium-rich rocks in the crystalline "granitic" crust has been a prominent process. The belt of Rockies through the shelf region seems to have affected the igneous suites very little—they are approximately the same in the Colorado Plateau, in the Wyoming and Colorado Rockies, and in the fairly stable area east of the Rockies in Montana. Their prolonged and complicated differentiation history bespeaks rather stable crustal conditions.

The Columbian River tholeiitic flood basalts are principally of Miocene age, center approximately in the great batholithic bulge, and have been fed upward through the batholithic complex (see Fig. 36.5).

The vent basalt field of the Snake River Valley and Malheur Plateau, is principally one of late Pliocene and Quaternary activity and occupies a downwarp around the south side of the Idaho batholith and, very approximately, along the south side of the great bulge which seems to be continued eastward into western Montana by the large Laramide plutons there. It is suggested that since the basalt came directly from the subcrust, and the downwarp is across the Laramide trends of the central Rockies, that the folding and thrusting is shallow and that the downwarp is due to movements in the subcrust. It must be noted, however, that a large part of the field lies on the batholithic belt and has been fed by basalt from the subcrust through the batholithic complex to the surface.

DISTRIBUTION OF PRIMARY MAGMAS

The map of Fig. 36.7 has been prepared to show the distribution of the different types of primary magma postulated to have given rise to the igneous rocks now displayed at the surface.

Two principal types of primary magma are postulated, namely, the

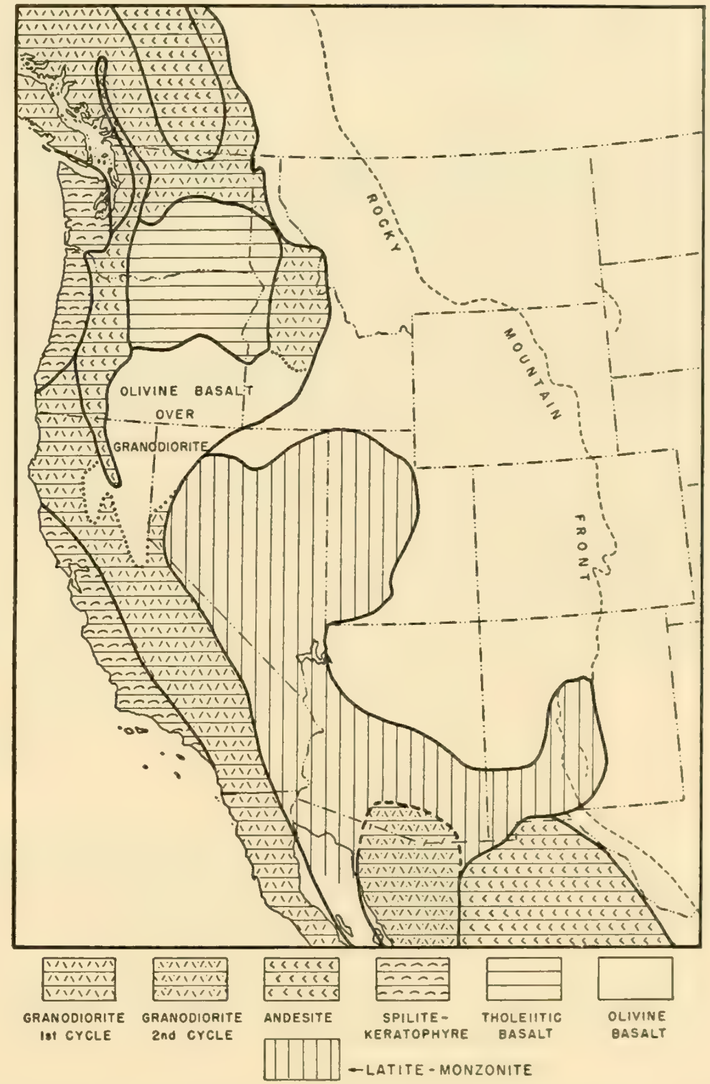


Fig. 36.7. Distribution of primary magmas. Batholiths of the second cycle occur in the Cascade andesite province. The spilite-keratophyre magmas of the pre-Nevadan batholithic time had about the same distribution as the batholiths. Andesite magma refers to the basalt-andesite association of the orogenic belts (post-batholithic).

basaltic and the granitic. The basaltic magma is of two classes, olivine and tholeiitic, with transitional varieties recognized. The granitic type ranges from tonolite to alaskite, and is considered to have originated in two slightly different ways. The origin of the primary magmas will be considered under later headings.

It is evident that both olivine basalt and tholeiitic basalt magmas have been conducted up through the granitic batholitic complex, and hence both varieties in large amounts can supersede the granodioritic magma in certain places. Smaller amounts of basaltic magma, probably all of the olivine variety, have made their way up through the crust in the province of the miogeosyncline or the latite-monzonite igneous province generally as a prelude or as a closing note to the main magnetic activity.

Olivine basalt is considered the primary magma of the alkalic and calc-alkalic provinces, although appreciable assimiliation and contamination of the magma has occurred. In a few places, considerable melting of short-lived roots may have occurred, and here, by definition, the primary magma would be granodioritic, quartz dioritic, or augite dioritic as locally identified. Even here, some basalt may have been mixed in.

The andesite and spilite-keratophyre provinces probably do not mark primary magma types. It is concluded on a later page too that they are fractional differentiates of primary basalt, probably of tholeiitic basalt, but because there is doubt about this conclusion they are shown separately (Fig. 36.7). The latite-monzonite province is concluded to be a primary magma province, although perhaps an unusal one.

37.

IGNEOUS AND TECTONIC PROVINCES OF WESTERN CANADA

GEOSYNCLINE

Eugeosyncline

The eugeosynclinal division of the Cordilleran geosyncline of western Canada and southeastern Alaska has been described in Chapter 6. Suffice it to say here that sediments of the eugeosynclinal type occur west of the Beltian geanticline of British Columbia. In southeastern Alaska strata of Ordovician, Silurian, Devonian, Permian, and Triassic age are laden with volcanics, whereas the lower and middle Paleozoic systems are not represented as far as known in the interior east of the Coast Range batholith. During the Carboniferous and Permian periods, however, the interior region subsided and great thicknesses of sedimentary and volcanic material accumulated.

Large areas in the Yukon and Interior plateaus are underlain by Triassic, Jurassic, and Lower Cretaceous formations made up of interbedded limestone, argillite, graywacke, conglomerate, tuff, breccia, and andesite flows. They have been invaded widely by great batholiths. Their original extent may have been approximately that of the batholithic belt of the map, Fig. 37.1, plus the areas shown as eugeosyncline. See also Fig. 17.13.

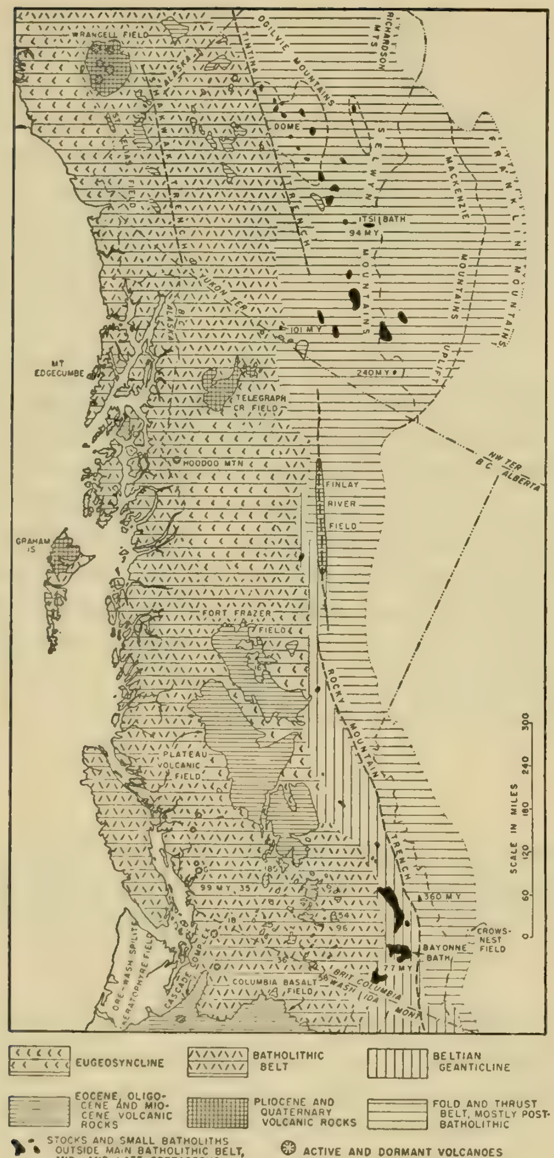
Miogeosyncline

The position of the Paleozoic miogeosyncline was that of the eastern Cordillera or more commonly referred to as the Canadian Rockies, east of the Rocky Mountain trench. The Cambrian and Ordovician strata are here particularly thick. The transition from the miogeosyncline to the Alberta shelf is probably a gradual one and lies under the Alberta basin. The miogeosyncline apparently dies out at about the Yukon Territory boundary on the north, and thence northwestward the eugeosyncline is transitional to the shelf. Thickness and lithologies in the Mackenzie and Selwyn Mountains are very poorly known, and therefore, also geosynclinal and shelf conditions cannot be very well discerned.

OROGENIES

The eugeosynclinal complex attests crustal unrest almost constantly. In places it was intense (see Chapter 5). Isoclinal folding with attendant low-grade metamorphism occurred in late Jurassic or early Cretaceous time to precede immediately the invasions of the numerous and large batholiths.

The Laramide belt embraces the eastern Cordillera, and possibly a wide region in Yukon Territory and in the western part of the Northwest Territories, including the Franklin, Mackenzie, Selwyn, and Richardson Mountains. The Mackenzie and Selwyn region is described as one of broad folds and a subordinate amount of faulting (Lord *et al.*, 1947). The folds are commonly arcuate and arranged *en echelon*. The Franklin



Mountains are probably part of this system, and the Richardson Mountains are also thought to be made up of folds with a northerly trend. The age in part from map relations appears to be pre- or Early Cretaceous, but elsewhere to be Late Cretaceous or Early Tertiary. As described in following paragraphs the age of one of the batholiths in the Selwyn Mountains is Mid-Cretaceous.

BELTIAN GEANTICLINE

A geanticline of Beltian strata emerges in southeastern British Columbia from the broad region of Beltian rocks in northwestern Montana, and extends northwestward in narrowing fashion almost to Yukon Territory. The Rocky Mountain trench lies along its east side for the most part. In northwestern Montana the deformed Beltian is on both sides of the trench zone, and in fact, forms the entire Rocky Mountain system there.

Broad areas of strata, probably equivalent to the Belt series occur in the Yukon, but since the whole region of outcrop is intruded by numerous batholiths, it is regarded as part of the Nevadan orogenic belt and, therefore, not shown as geanticlinal. The geanticline of Beltian strata in eastern British Columbia is probably one of Laramide orogeny. In Chapter 6 the geanticline is postulated to have developed as early as Cambrian time in the form of an arch separating the eugeosyncline from the miogeosyncline, but the main rise, evidently, was incident to folding and thrusting of the Laramide orogeny (see Chapter 20). The Beltian geanticline is somewhat of a parallel to the South American geanticlines.

BATHOLITHIC PROVINCE

The batholiths of British Columbia, Yukon Territory, and southeastern Alaska are arranged in two divisions or belts. Those on the west are described as follows by Lord *et al.* (1947):

Fig. 37.1. Major tectonic and igneous units of the Canadian Cordillera. G, Mt. Garibaldi. Numbers such as 102 m.y. are absolute ages in millions of years determined by the potassium-argon method (Follinsbee *et al.*, 1957).

The Coast batholith is the largest of the Mesozoic intrusions. It forms the core of the Coast Range and extends northwesterly about 1,100 miles from the northern part of the State of Washington to Yukon. Its width averages more than 50 miles and locally exceeds 125 miles. Flanking it for many miles on either side are smaller, related intrusive masses that, with the rocks of the main batholith, comprise what is commonly known as the Coast intrusions. In southern British Columbia the batholith curves towards the east and is linked with the presumably related Nelson batholith of Kootenay district by other intervening intrusions. The Coast intrusions range in composition from granite to gabbro, but are mainly of granodiorite and quartz diorite. The batholith is a composite of an unknown number of phases that were emplaced as successive irruptions over a long period of time, and, presumably, the numerous satellitic bodies are likewise of more than one age. The younger phases commonly show sharp intrusive contacts against older phases, and in many localities that batholithic rocks cut Lower Cretaceous sediments that contain pebbles of earlier batholithic rocks. It has been suggested that in northern British Columbia the more acid phases are most common towards the interior of the batholith. In the southern part of the province, however, the eastern intrusions, such as the Nelson batholith, are more acid and contain a greater proportion of granite than those nearer the coast.

A potassium-argon age determination of the Coast Range batholith near Vancouver is reported by Folinsbee *et al.* (1957) as 105 m.y. This is about Mid-Cretaceous on the Holmes scale.

The eastern belt of batholiths starts southwest of the south end of the Finlay River volcanic field at about Fort Frazer and Burns Lake (see *Geologic Map of Canada*, 1947) and extends northward to the Yukon and Alaska. In the Yukon the batholiths are so numerous that no marked division can be noted between those of the Coast Range belt and those of the eastern belt. Also at the south end of the eastern belt in the Burns Lake region numerous small plutons bridge the Coast Range batholith to the eastern batholiths. Between the south end and the northern cluster is the Cassiar–Omineca batholith, which is over 500 miles long but rather narrow. It is only partially explored and may not yet be completely unroofed by erosion.

These rocks commonly grade into one another and are not known to represent more than one continuous period of intrusion. Near Takala Lake the batholithic rocks cut Jurassic strata of early Upper Jurassic age. They also appear to have been the source of pebbles found in the early Upper Cretaceous conglomerate. Thus, so far as known, the main Cassiar–Omineca batholith is of Upper Jurassic or Lower Cretaceous age.

Intrusive, stock-like, tabular, and irregular bodies of serpentinized dunite, peridotite, pyroxenite, and gabbro are found in southern Yukon, in Dease Lake and Takla areas of northern and central British Columbia, and in Bridge River, Hope, Princeton, and other areas of the southwestern part of the province. The largest are more than 100 square miles in area, but most of them are much smaller. They are commonly considered to be early phases of the Mesozoic batholithic intrusions and to be of Jurassic age (Lord *et al.*, 1947).

A number of batholiths in the Selwyn–Mackenzie mountains salient seem to lie east of the main eastern belt, and might be thought of as Laramide. Yet a potassium-argon age determination by Folinsbee *et al.* (1957) on the Itsi batholith (Fig. 37.1) is 102 m.y. or Mid-Cretaceous. This is about the same age as the Coast Range batholith at Vancouver. It seems necessary to conclude that all the scattered batholiths in this salient are of the same age until further determinations are made.

In southeastern British Columbia the Bayonne batholith immediately east of Kootenay Lake has a potassium-argon age of 82 m.y. and is, therefore, about the same as the Boulder batholith of Montana to the southeast which is 87 m.y. old (Knopf, 1957). Both would be Late Cretaceous according to these dates and referable to the early Laramide phase of orogeny. This batholith and the one just north of it, are therefore placed in the Laramide belt of orogeny (see map, Fig. 37.1) and are considered late satellites of the Nelson batholith and the Idaho batholith.

POST-BATHOLITHIC VOLCANISM

In the Canadian Cordillera crustal movements occurred during the Tertiary from place to place along with considerable volcanism. The Tertiary disturbances, from what is known, consisted of faulting, tilting, and open folding. From Paleocene through Oligocene time sedimentary and volcanic rocks accumulated in numerous small basins unconformably on all older rocks. In Miocene and Pliocene time the major volcanism broke out and several large fields of nearly flat-lying units accumulated. The Miocene and Pliocene volcanics generally rest unconformably on the

older Tertiary volcanics which had been tilted and eroded. Some late Tertiary and Quaternary volcanic fields are also prominent. The volcanic rocks throughout are basalt and andesite lavas and related pyroclastics (Lord *et al.*, 1947).

The major areas of Tertiary volcanic rocks are as follows. In Yukon Territory and extending into Alaska irregular isolated areas occur which lie mainly within two northwesterly trending belts. The easterly of the two belts extends from near Carmacks, at the mouth of the Lewes River to and into Alaska, and is here called the Yukon field. The westerly belt lies along the northeast flank of the St. Elias Mountains and is here called the St. Elias field (see map, Fig. 37.1).

In northern British Columbia one area, in part of Quaternary age, extends north from Telegraph Creek on the Stikine River for about 80 miles (the Telegraph Creek field) and another floors the Rocky Mountain Trench for perhaps 150 miles along Finlay and Fox rivers (the Finlay River field). The latter consists of sediments and volcanics of late Oligocene and early Miocene age.

The largest field is in the southern interior of British Columbia where major accumulations in places several thousand feet thick extend 350 miles in a northwesterly direction and 150 miles in a northeasterly direction. These are flat-lying and commonly referred to as the plateau basalts. The accumulation is labeled "Plateau Volcanic Field" on the map. A large and associated field immediately to the northwest is here called the Fort Frazer. It consists in the Fort Frazer area of a lower series of upper Oligocene and lower Miocene sediments and volcanics dipping at angles up to 30 degrees and overlain unconformably by an upper series of nearly horizontal basalt, andesites, and other volcanic rocks about 2000 feet thick.

An informative Tertiary section is found in the Okanogan Valley close to the International Boundary. It is described as follows:

> Here the Springbrook formation, perhaps of Paleocene age, and composed of soils, alluvium, talus, stream and lake deposits, and tuff, rests on a pre-Tertiary rock surface of steep relief. These strata accumulated in the valleys and are overlain by and to some extent interlayered with the andesites, basalts, and pyroclastic rocks of the Marron formation, which buried the valleys and reached a thickness of more than 4,000 feet. The White Lake formation, consisting mainly of lake and stream deposits with coal, was deposited on the Marron strata from which most of their materials were derived. They are locally as steep as 65 degrees and 4,000 feet or more thick. Their age is probably late Eocene, but they may be somewhat younger. The White Lake strata are overlain unconformably by beds of more gently dipping andesitic breccia and agglomerate, which are succeeded upwards by agglomerate and conglomerate. The youngest conglomerate beds are horizontal and of pre-Pleistocene age (Lord *et al.*, 1947).

Volcanic activity occurred on a much reduced scale in the Quaternary period. In Yukon very young lavas occur and a loose, white volcanic ash is widespread which is at best only a few thousand years old. In the Telegraph Creek field Hoodoo Mountain on Iskut River may still be an active volcano. Recent lavas have been noted in several places along the coast.

RELATION OF VOLCANISM TO TECTONIC PROVINCES

Most all post-batholithic volcanism in the Canadian Cordillera is limited to the batholithic belt. Minor activity has occurred west of the major Coast Range batholith in the island archipelago, but the major activity was to the east of the island belt, and very approximately between the western zone of batholiths and the eastern zone. The fields are discontinuous and the volume of extruded rock is apparently not large. About one-tenth of the batholithic belt is covered.

The three trenches shown on the map, Fig. 37.1, are believed to be Tertiary grabens, but little is known about them. The Finlay River volcanic field fills the Rocky Mountain trench for a distance of about 180 miles. This is the only occurrence of volcanics in association with the trenches whose combined length in Canada is 3000 miles. It can be thought, therefore, that the association is accidental and not genetical. In South America, the trenches and volcanism seem more closely associated.

The row of stratovolcanoes of the Cascades extends into southwestern British Columbia. Pliocene, Pleistocene, and Recent activity is noted in

the Telegraph Creek field, at Mount Hoodoo and Mt. Edgecomb and in the Wrangell field, but this is an alignment only in the broadest sense, and separated by a 400-mile gap from the Cascade volcanoes.

The Rocky Mountain trench separates the Beltian geanticline from the Laramide Rockies in British Columbia. Its projection in Yukon, the Tintina trench, separates the batholithic belt from the Laramide (?) Rockies. If a geanticline is present there, it is very broad and is invaded by so many batholiths that it is represented as part of the batholithic belt.

The batholithic belt is about 400 miles wide at the international boundary and 300 miles wide at the Yukon–British Columbia boundary. At the Alaska border it is 200 miles wide. See Fig. 39.2.

38.

SPATIAL RELATIONS OF MAJOR TECTONO-IGNEOUS ELEMENTS AND THE ORIGIN OF MAGMAS

RELATION OF BATHOLITHIC BELT TO EUGEOSYNCLINE

The spatial relations of the major tectono-igneous elements of the western Cordillera of South and North America will now be summarized. The batholithic belt coincides almost exactly with the previous eugeosyncline. In places eastern segments of the eugeosyncline have not been invaded by the great batholiths, and in one known place, the batholiths have invaded the entire width of the eugeosyncline and also part of the miogeosyncline. This is in Idaho, western Montana, and southeastern British Columbia.

PREVIOUS OROGENY IN EUGEOSYNCLINE

In all parts of the eugeosyncline of South and North America, evidence of mid- or late Paleozoic orogeny is at hand, and especially in the Sierra Nevada of California and the Coast Ranges of southeastern Alaska we note a succession of orogenies of both Paleozoic and Mesozoic age. The isoclinal folding and development of slaty cleavage in the Jurassic Mariposa formation previous to the intrusion of the batholiths in the Sierra Nevada has clouded the effects of an orogeny of late Paleozoic age there, but the importance of the older orogeny is emphasized by a study of the eugeosyncline in South America where little note is made of the Late Jurassic or Early Cretaceous orogeny and commonly only the older Paleozoic one is recognized.

The Paleozoic orogeny in South America affected a belt from the present coast to the eastern mountain ranges in places. It shows today as the metamorphic basement of the Coast Ranges and as the metamorphic rocks in the anticlinoria 150 to 300 miles inland from the coast. The anticlinoria are in the eastern part of the Mesozoic eugeosyncline and in the western part of the miogeosyncline, and have formed during or later than the batholithic orogeny. The great width of the belt of dynamic metamorphism indicates orogeny of superior intensity long before the invasion of the great batholiths. The batholiths generally were emplaced in the oceanward margin of the metamorphic belt.

RELATION OF POST-BATHOLITHIC COMPRESSIONAL OROGENY TO GEOSYNCLINE AND SHELF

After the main batholithic intrusions, mainly in the eugeosyncline, strong folding and thrusting occurred in the miogeosyncline. In South America Cretaceous and Tertiary sediments have accumulated on the former transition region from miogeosyncline to shelf, and thus the relation to folding is obscured. Aside from the Pampean Ranges it appears that no conspicuous orogeny has occurred in the shelf. The width of the belt of folding ranges from 50 to 300 miles.

Post-batholithic folding in Mexico is extensive. Laramide folding from

the Nevadan batholiths to the east front of the Sierra Madre Oriental forms a belt up to 450 miles wide. Its relation to Paleozoic sedimentary basins is largely unknown, but it embraces the Cretaceous eugeosyncline and miogeosyncline.

In the United States folding and thrusting extend through the miogeosyncline. See Fig. 36.6. Basins and asymmetrical anticlinal uplifts expressive of significant vertical components of force are common in the shelf. The maximum width of the belt of folding in the Paleozoic miogeosyncline is about 300 miles, and the front of the belt of deformation in the shelf is another 400 miles farther east at its most easterly point. As in South America rather thick Mesozoic sedimentary sequences have accumulated along the Paleozoic miogeosyncline and shelf transition zone and also in places over the shelf. The shelf in the United States was also the site of building of the Ancestral Rockies in late Paleozoic time.

The belt of Laramide folding in British Columbia and Alberta is intense and about 100 miles wide. It is confined to the Paleozoic miogeosyncline and the western part of the Mesozoic basins over the miogeosyncline and shelf. Farther north, the folding is less intense but has a maximum width of about 300 miles. It spreads here mostly over a Paleozoic cratonic basin and an overlying Cretaceous basin, and the force component is vertical.

RELATION OF POST-BATHOLITHIC VOLCANICS TO BATHOLITHIC BELT

In Chile south of Antofagasta and Argentina the post-batholithic volcanics are chiefly on the deformed eugeosyncline, miogeosyncline, and shelf east of the batholithic belt. The maximum width of the general belt of extrusive rocks is 300 miles. In northern Chile and southern Peru the belt of volcanic deposits is partly on the batholithic belt, and spans across the narrow eugeosyncline to the deformed miogeosyncline. Only a small amount of extrusive rock is found as far east as the anticlinorium. The main belt of volcanism is about 150 miles wide.

The belt of volcanic deposits in Ecuador and Columbia spreads across the boundary of the batholithic belt and the anticlinorium. This volcanic field is narrow in comparison with the others, and does not exceed 50 miles unless some of the stratovolcanoes are considered part of it, in which case the belt reaches 80 miles in width.

There are two major volcanic fields in Mexico. The larger of the two, the Sierra Madre Occidental, rests on the deformed Cretaceous eugeosyncline and in places over the miogeosyncline, a considerable distance east of the Nevadan batholithic belt. The smaller field, and the more recent, rests on the batholithic complex in the southern part of Baja California. The great Sierra Madre Occidental field is 200 to 300 miles wide. The smaller field in Baja California is 30 to 60 miles wide.

The volcanic fields in western Canada lie almost entirely within the batholithic belt. The largest accumulations, in British Columbia, are 150 miles in width.

The volcanic fields of the western United States are broad and varied. Where the batholithic belt is narrow in southern and central California, volcanic deposits are few, but all through the broad miogeosyncline 150 to 300 miles wide, they are extensive. They also occur in considerable quantity in scattered fields in the shelf to the east which in part has been moderately deformed in post-batholithic time. Volcanic eruptions in Colorado are 500 miles east of the miogeosyncline, and the Black Hills igneous rocks are 350 miles east of the miogeosyncline.

In northwestern United States, where the Nevadan belt is very wide, the great basalt fields occur. The tholeiitic (Columbia River) basalt field is entirely on the batholithic belt and is nearly 300 miles wide. The olivine vent basalt field is mostly on the batholithic complex but extends eastward over the miogeosyncline to the shelf. These two large basalt fields are exceptional to all other fields in the western Cordillera of South and North America, and seem related to the great batholithic bulge at the intersection of two Nevadan orogenic arcs.

RELATION OF POST-BATHOLITHIC VOLCANIC FIELDS TO STRATOVOLCANOES

The three rows of stratovolcanoes of the South American Cordillera are closely related to the orogenic andesite complexes. The southern row, south of Santiago, however, is not accompanied by voluminous fields; the

volcanoes for the most part stand as isolated piles having been fed by conduits through the batholithic complex and deformed eugeosyncline.

The stratovolcanoes of the southern Cascades of Oregon and Washington have been built on an older andesite complex but in the northern Cascades of northern Washington and southwestern British Columbia they stand as isolated cones fed by conduits through the batholithic complex.

The stratovolcanoes of southern Mexico are built on an extensive older volcanic field and are part also of extensive fields evidently as young as the volcanoes themselves. The belt of stratovolcanoes seems to lie on the inner margin of the metamorphic belt and also partly on the deformed adjacent geosyncline. See Chapter 43.

POST-BATHOLITHIC VOLCANICS TO TRENCHES

Trenches are of two kinds, the submarine deeps marginal to the continents and the fault-depressed zones generally within the batholithic complex or separating it from the anticlinoria. The volcanics are of the stratovolcanic and basalt-andesitic field types.

The fault-depressed trenches are the sites of both basalt-andesite complexes and stratvolcanoes. The stratovolcanoes occur in both the depressed blocks and on the adjacent upraised blocks. In Chile south of Santiago the cones are chiefly on the east side of the depressed zone. In northern Chile and southern Peru the volcanic deposits may have filled depressed blocks, with extensively faulted regions both east and west of the volcanic accumulations. In northern Peru, Ecuador, and southern Columbia, the volcanic deposits have filled a long graben-like depression, and stratovolcanoes are present within the depression and on its marginal uplifted blocks, particularly on its eastern block.

The fault-depressed blocks of the South American Cordillera are generally described as compressional structures, bounded on one side or both by uplifted overthrust blocks. The southern Mexico stratovolcanic province is probably bounded on the south by a block-faulted region, but little is known of the structures there. The great Sierra Madre Occidental field is broken and bounded on the west by a thrust-faulted zone and then by the major depressed zone of the Gulf of California. The depressed zone here is postulated to be due to the drift of Baja California away from the continent and to the northwest, in connection with the San Andreas fault movements of California. A young volcanic field exists on the west or outer side of the depressed zone.

In western Canada the andesite complex is west of the depressed zone, here the Rocky Mountain Trench, which separates the Nevadan complex and geanticline from the deformed miogeosyncline.

In the United States the relations are very complicated, and comparisons with the South American and Canadian can only be imagined. The broad Basin and Range province would be the fault-depressed belt, which in Nevada and Utah is superposed on the eastern side of the eugeosyncline and across the entire miogeosyncline. It is replete with volcanics but not of the basalt-andesite complex, but rather of the monozonite-latite clan. A zone of particularly conspicuous trenches (graben and horst blocks) make up the eastern side of the Basin and Range province and these extend northward through Idaho and western Montana into British Columbia. Relatively minor volcanic activity is noted in the zone of trenches from the High Plateaus field of south-central Utah to the Finlay River field of northern British Columbia.

A spatial coincidence of the stratovolcanoes of South America to the offshore, submarine trenches is immediately conspicuous, but in detail we may note first, the Chilean row south of Santiago extends southward beyond the limits of the submarine trench and second, the trench is continuous but the stratovolcanoes occur in three separate rows or segments.

The Central American Trench lies opposite the stratovolcanoes of southern Mexico and also the active and dormant volcanoes of Central America, but the trench, as an ocean-floor phenomenon, does not continue northward where the major volcanic complex of Mexico occurs. The submarine trench coincides well with recent volcanic activity but not with the older activity.

The Cascade andesite complex and row of stratovolcanoes is not complemented by a submarine trench. It may, therefore, be concluded that a submarine trench is not a necessary accompaniment of a row of adjacent stratovolcanoes; one may exist without the other, but their coincidence spatially is more likely than not.

It may also be concluded that trenches within and east of the batholithic belt are nearly everywhere present along the entire Cordillera of North and South America, and that in places volcanism seems fairly well localized to the trench or immediately adjacent to it. Extensive volcanism occurs in Mexico, however, on either side of and at a considerable distance from the depressed zone.

RELATION OF ANTICLINORIA TO OTHER ELEMENTS

Anticlinoria of Precambrian or metamorphic Paleozoic rock occur parallel to and on the inside of the batholithic belt. These generally elevated areas are encompassed in the belt of post-batholithic folding and in places are separated from the batholithic zone by the fault-depressed zone. The anticlinoria range in width from 50 to 150 miles. They are present in the more typical South American and Canadian Cordillera but not present in the atypical United States Cordillera. They lie generally east of the major volcanic fields, although some volcanics occur on them and even east of them.

ORIGIN OF MAGMAS

Physical Considerations

Crustal Structure. The crust forming the continental masses according to seismic information (Tatel and Tuve, 1955), has a general thickness of 28 to 35 kilometers, but interpretations as low as 20 kilometers in coastal California and as great as 65 kilometers under the Sierra Nevada and 72 kilometers under the eastern Great Basin are given. An abrupt change in seismic velocities at the base marks the Mohorovicic discontinuity which is believed to be world-wide.

The upper layer of the crust which has low velocity is called the granitic crust, silicic crust, or sial, and the lower, the basaltic crust, gabbroic crust, sima, or subcrust. Tatel and Tuve concluded that the two are probably transitional, but others have postulated distinct layers locally of intermediate velocity and of different relative thicknesses.

The silicic layer consists of the lighter rock-forming silicates and is high in Si and Al, and the basaltic layer, as the name implies, consists of the darker and heavier silicates and is lower in Si and Al and higher in Fe and Mg.

The floor of the oceans, beneath a thin veneer of lava flows and unconsolidated sediments, consists of a basaltic layer 5 to 10 kilometers thick, which overlies the mantle. Extensive volcanic accumulations, also composed mostly of basalt, rest on the basaltic crust in many places.

The outer part of the mantle down to a depth of several hundred kilometers is crystalline. It consists mainly of dense silicates of Mg and Fe, prominent among which is olivine, and is often referred to as peridotitic (Turner and Verhoogen, 1951), but many be eclogite, a high-density phase of gabbro (Kennedy, 1960).

Geothermal Gradient and Melting Points. Measurements in mines and wells indicate that the earth temperature increases with depth at a rate of approximately 30°C per kilometer. Gradients as low as 7°C per kilometer and as high as 50°C per kilometer are known but are exceptional. According to Turner and Verhoogen (1951) the temperature at the base of the crust, say at 40 kilometers, is 500 to 600°C, at 100 kilometers 800 to 900°C, and at 2900 kilometers 1500°C.

Magmas erupted from volcanoes have been found to have temperatures as high as 1000 to 1200°C, and the melting temperature at the surface of basalt of about 1000°C is in this general high-temperature range. But such a temperature is not normal to the rocks at the 40-kilometer depth. Consequently, according to Turner and Verhoogen, either the magma originates by fractional melting of deep-seated earth material of peridotitic or eclogitic composition, or it is the result of melting of shallower rocks in place under temperatures temporarily raised far above the average temperature normally prevailing at that depth.

Temporary and local increase in temperature within the crust or outer shell of the mantle might be developed in three ways: (1) by the blanketing effect of a thick sediment-filled basin; (2) local radiogenic heat; and (3) frictional heat due to diastrophism. Turner and Verhoogen conclude that the blanketing effect of sediments 10 kilometers thick would result in an increase in temperature of less than 200 or 300°C in the crustal rocks beneath. Regarding radiogenic heat in the outer mantle shell, they believe that this could result in cyclical convective overturn, and that the temperature of the crust immediately above would be raised appreciably

above its normal value by conduction with each fresh overturn of the convection cell. This should correlate with intermittent magma generation, and possibly the development of the 7.5-kilometer-per-second seismic velocity layer (Chapter 31).

Heat generated by crustal deformation has been held very significant by some, and the epigram "Diastrophism is the mother of volcanism" is commonly recited; yet widespread, and in places voluminous, magmatic activity has occurred in regions of crustal stability. As concluded in the chapters on the Rocky Mountains, igneous activity may be the cause of the diastrophism. There must be a real tie between the batholiths of the Nevadan belt and crustal deformation, and also between the later andesitic-basalt eruptions and diastrophism. In reference to the great mantle fault, shown in Fig. 38.3. Benioff suggests that considerable heat is generated in the aftershocks which are a manifestation of creep strain in the rocks, and that this heat may be sufficient for the apparently related volcanic activity. He says,

> A rough idea of the magnitude of energy released, say per year, by the aftershock sequences in a region on one side of the fault can be obtained by taking one fourth of the energy released in the same time by seismic waves in the principal earthquakes. Thus, in the case of South America, the shallow and intermediate earthquake sequences each liberate approximately 4×10^{23} ergs per year. Thus roughly 10^{24} ergs per year is being released in the fault blocks. The writer has no knowledge of the amount of energy per year required to maintain the South American system of volcanoes, and consequently it is not possible to say whether or not the energy requirements are met on this hypothesis. Moreover there must be a large time lag between the liberation of heat in the depths and its appearance in the form of volcano output. Thus the present rate of volcanic energy release should be equated to a phase of seismic-heat generation which occurred long ago, rather than to the present rate.

The problem of the origin of magma is not one of quantity of energy according to Turner and Verhoogen (1951):

> . . . radiogenic heat in the earth seems to be ample to account for all geologic (including igneous) phenomena, but what must still be sought is some process which will concentrate this energy locally, and raise the temperature sufficiently, at points of concentration. Igneous activity itself testifies to the operation of some such process. But its precise nature remains an unsolved problem.

Primary Magmas

Definition. Primary magma, by definition, originates by partial or complete fusion in great volume of pre-existing rock. It is conceivable that some igneous bodies have come from a primitive liquid still existing from an early stage in the earth's history but no satisfactory evidence for such has been recognized (Turner and Verhoogen, 1951). The modification of a primary magma results in *derivative* magmas.

> Criteria by which a primary magma may be recognized as such are somewhat vague. Probably the most satisfactory is a pronounced tendency for the magma to appear repeatedly throughout geologic time, in great quantities and in extensive individual bodies (lava floods, batholiths, lopolithic sheets, etc.), over large sectors of the earth's crust. A further criterion is predominance of corresponding rocks within one or more rock associations, the other members of which could have been derived from the primary magma by accepted modifying processes—differentiation, assimilative processes, etc.
>
> Conversely there is a tendency to regard magmas as belonging to the derivative class when they occur habitually in small quantities, when they are constantly found in association with a magma conventionally considered as primary, and when derivation from the latter can be explained in terms of accepted modifying processes (Turner and Verhoogen, 1951).

Classification. There is general agreement that two broad primary magma families exist, namely, granitic and basaltic. By granitic is meant the common associates, granodiorite, quartz monzonite, and granite, and perhaps tonalite, diorite, and others closely akin which in places occur in great volume. Extrusive andesite is regarded by Waters (1955) as a primary magma, but this is questionable. Its relation to the granitic group will be discussed later.

The basalt family is made up of two main varieties, namely, olivine and tholeiitic. Gradational varieties are common.

Magmas of the Alkalic Igneous Province

Prevalence of Olivine Basalt as Primary Magma. Under a previous heading in connection with Fig. 36.6 it was concluded that the exposed igneous rocks of the alkalic igneous province of the western United States were derived from a primary olivine basalt magma. It was also postulated that the surficial intrusions and extrusions come from megasills in the sur-

ficial granitic crust where various amounts of assimilation have occurred. Of course, magmas intrusive into shallow sedimentary sections have affected the overlying beds such as over laccoliths and bysmaliths, and even over and around stocks in places, but these structures could not be related to the origin of the magma.

The Laramide structures of the alkalic province appear not to have roots as previously suggested, and one of the most intriguing geophysical studies is the seismic charting of the velocity layers in pursuit of this problem, and also the source of magmas there.

Seismic Evidence of Crustal Structure. Tatel and Tuve (1955) report the base of the basaltic layer (Mohorovicic discontinuity) under the Colorado Plateau (part of the alkalic province) at the shallow depth of 30 kilometers. This was surprising because from isostatic considerations the high plateau should have been supported by a crust some 50 to 70 kilometers thick (50 to 70 kilometers to the M discontinuity). From this and other data they conclude neither the Airy nor Pratt concepts of crustal structure hold. Gravity observations indicate a continent over which there is isostatic compensation, and therefore, they conclude that the outer mantle below the crust has density variations (see Chapter 31). Thus, in turning to the outer mantle for causes of vertical movements, a column, perhaps one to several hundred kilometers thick (or long) may be involved, and if so, only small changes are necessary to elevate the plateau 5000 to 8000 feet.

Basaltic Magma from the Mantle. Basalt magma can originate (1) by fractional melting of deep-seated earth material of different composition, (2) by complete melting of a deep-seated rock of the same composition, or (3) by complete melting of shallower rocks temporarily raised far above the average temperature normally prevailing at that depth. If surface basalts come from the mantle the process has been postulated to be one of partial melting of a basic rock of the composition of stony meteorites (peridotitic), or of melting of eclogite, a heavy crystalline rock of basaltic composition. Also the subcrust, presumably of basaltic composition, might melt in places to form a basaltic magma. The primary olivine basalt of the Laramide Rockies is believed by the writer to have come from the mantle and the following reasons are given.

1. The seismic evidence of layering, and the consequent interpretation of gravity measurements indicate that rock density changes must occur in the mantle. The Laramide structural province generally lacks roots and is underlain by the thick 7.5 layer. Therefore, density changes in the mantle are almost the sole explanation of isostatic adjustments. This suggests that magmatic processes may be occurring there.

2. The two primary basaltic magmas, tholeiitic and olivine, are best explained as coming from the mantle. See discussion under a later heading, "Tholeiitic Magma."

3. The heat necessary for local partial melting of the upper part of the mantle in places may adequately be provided by underlying convective overturn, or possibly by generation along faults deep in the mantle.

If partial melting of spots in the outer mantle shell is postulated as the source of primary basaltic magma, a series of consequences must be envisaged, which, if the theory is correct, must fit the pattern of structures and igneous intrusions and extrusions through time and space as well as geophysical observations and analyses. First, there are the considerations of expansion. Since the Colorado Plateau is some 2 kilometers above sea level and has no roots to buoy it up, we can think of expansion of the mantle beneath to have raised the crust. In melting, a volume increase of 11.2 percent occurs, and if 6 percent of a 300-kilometer column should melt, the crust would be elevated 2 kilometers. There would also be expansion in the solid state of the column over the convection cell as heat is conducted upward. The problem is complicated and will require the attention of experts, but evidently the amount of solid expansion is considerable. The concept of rise of basaltic magma from the mantle is attractive because it presents a plausible theory of the origin of the basaltic subcrust. Instead of a primitive basic differentiate of a more silicic melt from above, the subcrust would be the result of additions through time from below. This view would hold for the basaltic substratum under the continents, but for the ocean floors we would need to think of early outpourings, and after sufficient accumulation, increasing amounts of magma from below to build up the basaltic layer.

In connection with the arrest of basaltic magma from the mantle in the subcrust we may think of uplifts like the Black Hills, Big Horn Moun-

tains, Uinta Mountains, and San Rafael Swell as results. These elongate uplifts have lengths of 75 to 150 kilometers and widths of 40 to 75 kilometers. Giant-sized laccolithic intrusions in the subcrust of similar horizontal dimensions could have originally arched and upfaulted the structures, whose structural relief would thereafter have been augmented by sediment transfer to adjacent basins and gravitational adjustments. A giant-sized laccolith would need to be perhaps only 1 kilometer thick to result in a final structural relief at the surface of perhaps 3 or 4 kilometers. See Fig. 36.4.

Still other considerations of the theory of primary basalt magma generation in the outer mantle remain. They are in the fields of gravity and seismicity. Dr. Kenneth L. Cook's reactions to the gravity problems are as follows. If the outer shell of the mantle should expand and elevate the crust, say of the Colorado Plateau, 2 kilometers, isostatic anomalies in the order of 10 to 20 milligrams would probably occur, but effects of local surficial density variations might mask the overall isostatic anomaly picture to the extent that it would be unrecognizable. The problem is fairly complex. At least the concept of partial melting and expension of the outer mantle under regions as large as the Colorado Plateau does not run afoul of any gravity observations or interpretations that he could see off hand.

Dr. Joseph W. Berg's reactions to the seismic problems are as follows. Melting of 5 percent of a certain column of the mantle in a disperse system of some kind would lower earth wave velocities, but not any more than the observed range of velocities interpreted from seismic records in the upper mantle or lower crust. As far as he could see, the concept of partial melting of parts of the upper mantle 50 to 200 miles across is not contrary to any seismologic analysis.

From the above considerations it is concluded that under the Laramide systems of the alkalic igneous province olivine basalt was generated by partial fusion of the upper mantle and rose to the subcrust where it was intruded, probably in giant sill bodies; only minor amounts escaped up through the silicic crust to the surface. Large bodies of the molten basalt lay directly under the silicic crust, affected some melting and assimilation, and by various routes of fractional crystallization, mixing, and sieving, the contaminated primary magmas bore in small amounts the unusual alkalic and calc-alkalic suites of the Colorado Plateau, Wyoming, and Montana.

Magmas of the Nevadan Systems

The conclusion has been reached on prevous pages that the batholithic masses of the Nevadan belt represent such an enormous bulk of quartz-monzonitic and granodioritic material that it is impossible to conceive of their derivation from a basaltic parent by fractional crystallization, and, providing they were once mobile, we are forced to conclude that they represent a primary acid magma. Further, the primary magma originated by the melting of a part of the silicic crust in a master belt of orogeny along the continental margin. The conventional concept involves a thickened crust whose roots melt. The thickness of the silicic layer under the Sierra Nevada is now about 20 kilometers and about 25 kilometers in north-central Utah, but possibly before melting and isostatic adjustments, the crust there was much thicker. See Fig. 38.1. The basaltic subcrust seems about as thick as the silicic crust under the Sierra Nevada, but if upward adjustment has occurred after orogeny then the silicic crust would have been thinned by erosion, as well as viscous flow, and this consideration points to a previous much thicker silicic crust.

The theory of origin of primary basalt, therefore, contrasts sharply with that of primary granodiorite; the first by partial fusion of the upper mantle shell and upward migration to the subcrust and crust, and the second by fusion of large masses of the lower part of the thickened silicic crust in belts of master orogeny.

The above discussion is in the manner of those who believe that the great batholiths were emplaced by mobile magma, but there are many authorities who believe that the batholithic rock formed in place by a transformation of previously existing rock. Strong evidence is presented to support this point of view, namely that of granitization. For a review of the evidence see Gilluly (1948). One's point of view changes radically in considering crustal layering, roots, and intrusion space problems when convinced that granitization is the process at hand. It will be commented on later under the headings of andesite magmas and quartz latite magmas.

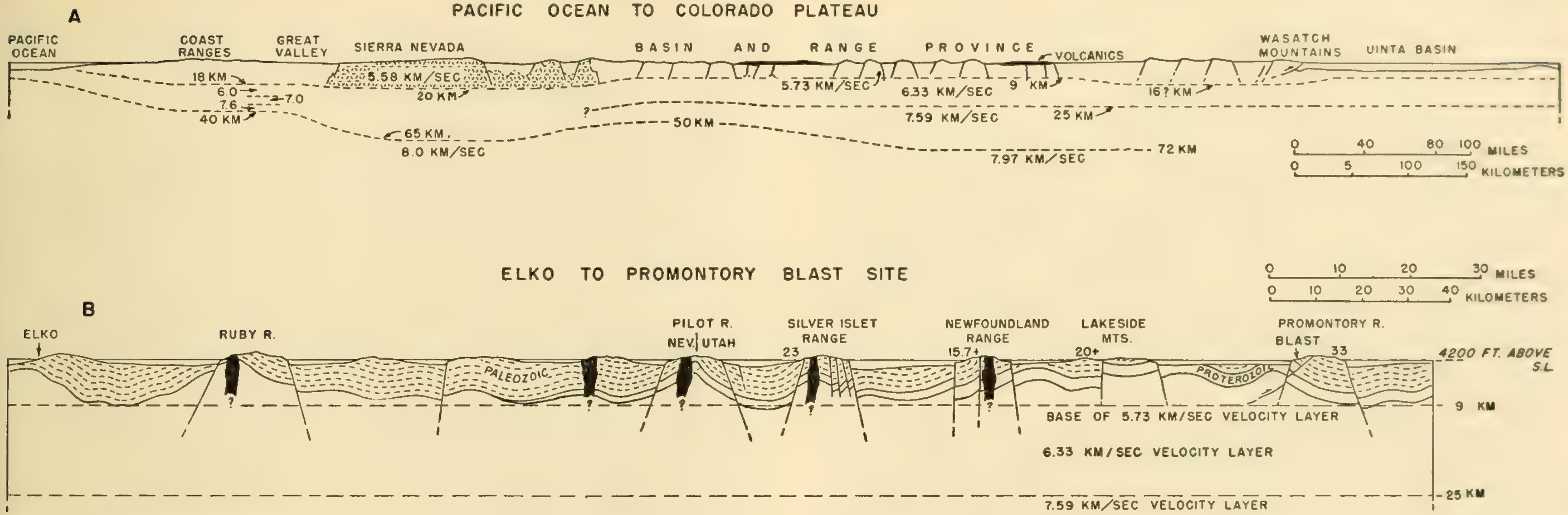


Fig. 38.1. Postulated seismic layering in relation to geologic structures at the surface of the western Cordillera. Velocities in Wasatch and Great Basin from Berg *et al.* (1960); for western Great Basin, Press (1960) and in the Sierra Nevada, Gutenberg (1943).

Tholeiitic Magma

The occurrence of theoleiitic basalt in eugeosynclines, over a wide area of the older Nevadan batholithic complex, in the Triassic fault basins of the Appalachian Mountains systems, in the Paraná basin of the stable shield area of Brazil, and in the Hawaiian Islands of the Pacific basin indicates that no tectonic setting has a monopoly on the magma. Previous considerations have shown that there is a fairly complete range from olivine basalt to tholeiitic basalt, if world-wide examples are tabulated together, but in local occurrences the spread is usually small, and separate igneous provinces may be recognized.

Convincing petrographic evidence for the origin of basalt by partial melting of the mantle comes from olivine-rich nodules in basalts. These are concluded by Ross *et al.* (1954) to have been xenoliths derived from the peridotitic mantle, and studies by Kuno *et al.* (1957) lead them to the same theory. Such xenoliths are known from about sixty localities scattered throughout the oceanic as well as continental regions. Kuno concludes that most of the olivine-rich nodules occur in alkali olivine basalt and allied rocks such as andesine andesite, nepheline or leucite basalt, and basanite and limburgite. Only a few doubtful examples in tholeiites are known. This distribution was first thought to signify that only the alkali olivine basalt comes from the mantle, and that the tholeiitic originates from bodily melting of the basaltic subcrust, but when it was realized that the Moho discontinuity under the Hawaiian Islands is only 5 kilometers deep, it was concluded that temperatures could not become high enough in the basaltic subcrust to cause melting. Kuno concluded, consequently, that both basalt types originate by partial melting of the upper mantle.

From extensive petrographic and chemical studies, particularly in

Hawaii, Kuno *et al.* (1957) conclude that neither basalt magma may be derived from the other by fractional crystallization, or in other words, that neither is the parent of the other. This emphasizes again the origin by partial melting of the peridotitic mantle.

Two possible reasons may be presented for the origin first of one magma and then of the other, or of one magma in one place, and the other in another place. The first is based on the assumption that the mantle is slightly heterogeneous in composition and that by partial melting of one part an olivine-rich basalt will be produced and by partial melting of another part with a slight difference in composition a tholeiitic basalt will result. Since these variations in the mantle are not tied to the tectonic divisions of the continental or oceanic crust in any way at present recognizable, it would thus be apparent why either variety of basalt rises in most any tectonic setting.

Kuno proposes a second possibility, namely that the mantle is of uniform composition and that different pressures will cause slightly different melting of the peridotite. In the Japanese Archipelago he suggests that the parental tholeiite magma is produced by partial melting of the peridotite layer (mantle) at depths shallower than 200 kilometers, and that the parental alkali olivine basalt magma is produced by partial melting at depths greater than 200 kilometers (Kuno, 1959).

Basalt-Andesite Assemblages of the Eugeosyncline and Orogenic Belts

On previous pages we have seen that the igneous rocks, so abundant in the stratified sequences of eugeosynclines are principally andesite and tholeiitic basalt. Spilites are common, and according to Waters (1955) they are tholeiitic basalt altered mostly by rising hydrothermal solutions but in part by sea water in connection with submarine flows. The albite may also be added in subsequent metamorphism, but in any case, they do not therefore add to the problem of the origin of the association of tholeiitic basalt and andesite. Keratophyres bear much the same relation to andesite as the spilites do to basalt and, hence, likewise do not pose an additional problem in the nature and origin of the primary magma. Olivine basalt is reported in places in the eugeosyncilinal assemblage but information on its relative volume and distribution is not well at hand; nevertheless it seems that provision should be made for its presence in the eugeosyncline in any theory of origin devised of the igneous complex there. Certain acid varieties are present in small amounts and are undoubtedly derivatives of the others.

The association of basalt and andesite in the volcanic fields of post-batholithic age has been elaborated on in previous pages. Reference to the cross sections of Fig. 34.5 and the map of Fig. 36.5 indicates that the site of most extensive occurrence is on the deformed belt immediately inside the batholithic belt toward the continent, which embraces parts of the older eugeosyncline not metamorphosed and invaded by the batholiths, and most or all of the miogeosyncline. In the United States where a belt of Laramide deformation is beyond the miogeosyncline in the shelf region, andesites also occur. About half the bulk of the San Juan field in Colorado is andesite, the other rhyolite, with basalt subordinate, so it is evident that somewhat different conditions apply there. The broad Great Basin region of the western United States is also unusual in relation to the general composition of the great western Cordillera of the Americas and will need special consideration.

The Cascade volcanic complex of Oregon and Washington should be mentioned in regard to post-batholithic activity because of significant associations there. It, however, is not a parallel with apparent normal conditions in the Cordillera, because it is a local field on the batholithic belt and also probably on the oceanward side of the batholithic belt built as part of a new continental margin. The older Cascade complex is more variable than that of the younger stratovolcanoes and according to Waters is a tholeiitic-andesite assemblage with some olivine basalt present whereas the younger is an olivine basalt-basaltic andesite assemblage.

Andesite is also found on Hawaii, in an olivine basalt, ocean basin assemblage. In connection with this occurrence and with the transitional nature of basalt and andesite the following quotation from Williams *et al.* (1954) is significant.

> Olivine-bearing andesites. These are widespread on oceanic volcanoes, like those of the Hawaiian Islands, and in orogenic belts of the continents. Indeed they probably predominate among the Tertiary and Quaternary lavas of the Circum-Pacific belt. Many of them lie so close to the boundary between

andesite and basalt that only chemical analyses serve adequately to classify them; in default of analyses, these borderline lavas are sometimes spoken of as "basaltic andesites." Olivine and labradorite may be their principal minerals, yet their silica content and the presence of normative quartz relate them to the andesite family.

Another variety is pyroxene andesite which according to the above authors is especially common on large composite volcanoes in the orogenic belts. Still others are hornblende and biotite andesites. These generally form thick short flows, steep-sided domical protrusions, or intrusive plugs and dikes, and are generally more siliceous and alkaline, and graded into dacites and trachyandesites.

With the above observations about the tectonic distribution and petrologic relations of andesite in mind, we must recognize four possibilities of origin: (1) a rock of andesitic composition melting completely and furnishing a primary andesitic magma; (2) a more basic rock melting and partially freeing a liquid of andesitic composition; (3) a granodiorite-granite primary magma mixing with a primary basaltic magma to form an andesitic magma; or (4) formation of an andesite by some variation of fractional crystallization.

The fourth category has two variations according to Turner and Verhoogen (1951). By fractional crystallization tholeiitic basalt may yield andesite; and a primary granodiorite-granite magma may yield andesite as a basic differentiate. The fractional crystallization of an alkali olivine basalt, according to Kuno (1959), could not result in an andesite, but instead various rocks like trachybasalt, nepheline basalt, trachyte, phonolite, or syenite would form. A calc-alkalic olivine basalt, however, can differentiate to an andesite (Kuno, 1959). The andesites in the San Juan field are regarded by Larson and Cross, with various mixings and contaminants, to have come from an olivine basalt. It is concluded that both tholeiitic and olivine basalt can give rise by differentiation to one variety or another of andesite.

For the small quantities of oceanic andesite the process of fractional crystallization from a tholeiitic basalt seems the most likely origin. This theory necessitates the presence of tholeiitic basalt in an olivine basalt province, but fortunately tholeiitic basalt is present in some oceanic islands.

As to the mixing of primary granodioritic magma with primary basalt magma to yield andesite magma, the process conceivably could occur in the root region of the batholithic belt and would involve the rise of primary basalt from below, according to the theory proposed on previous pages for the origin of either primary olivine or tholeiitic basalt. This could probably produce the necessary large volumes of andesitic magma necessary and also the transitional varieties from the two primary types. Mixing is not possible if the batholiths form by granitization.

In connection with the concept of roots of the batholithic belt it does not seem logical to think of them at one time melting to form a magma of granodioritic and granitic composition and then later melting to form one of andesitic composition. This is probably a good argument to the effect that the batholiths formed in place, and have nothing to do with roots. If it is accepted that the batholiths form by granitization, then it seems possible that roots, if they exist, could melt to form the andesites. It has been suggested, also, that the eugeosynclinal sequence of graywacke, argillite, and basic volcanics, if melted in bulk, would form a magma of andesitic composition. Inspection of the maps of South America, Figs. 34.1 and 34.2, will reveal that the large basalt-andesite complexes spread about equally over the eugeosyncline and miogeosyncline, so that the rocks in the eugeosyncline do not seem to have a direct bearing on the origin of the andesitic magma.

By elimination then, and with a bias for the magmatists, we arrive at the conclusion that the andesites are differentiation products of basaltic magmas, which vary in composition themselves between olivine and tholeiitic. The andesites in the alkalic and calc-alkalic provinces (tectonically the shelf, partly deformed in the Laramide orogeny) are probably a different breed from those of the deformed eugeosyncline and miogeosyncline, and have come about through an eventful history of mixing of differentiating magmas, and by appreciable assimilation of high calcic and alkalic rocks of the silicic crust. The andesites of the eugeosyncline and post-batholithic orogenic belts are only a step away from the basalts, the more acidic differentiates are centainly in the small minority, and the volumes of andesites and basalts are great, and the succession of flows and repetition in space monotonous.

If the andesites of the basalt-andesite complexes of the eugeosynclines and orogenic belts are differentiates of basaltic magma, then large volumes of the rising basalt from the upper mantle are trapped or arrested in the basaltic subcrust, where partial fractional crystallization and the development of andesitic and basalt-andesitic magmas takes place. Then as fissures in the overlying crust come into existence various magma pools are tapped, which may be basaltic through transitional phases to andesitic or even in rare instances, dacitic, and the surficial basalt-andesite complexes are extruded. In the voluminous outpourings andesite is commonly the most acidic rock produced, and so it would appear that the arrested bodies or reservoirs of magma in the subcrust are sill-like and not very thick, otherwise if in large bodies of several kilometers in vertical dimensions more varied and more silicic magmas might result.

Why the restriction to the orogenic belts and the eugeosynclines? The eugeosynclines are essentially orogenic belts themselves, but probably without appreciable roots until involved in the climactic batholithic orogeny. The blanketing sediments of the geosyncline result in the rise of temperature in the underlying crust and upper mantle, and hence may be thought of as bringing on the subcrustal igneous cycle. However, basalt has risen under the shelf of the stable region in considerable amounts without a thick, widespread, sedimentary blanket. The miogeosyncline developed irregularly with basins and arches, and these from previous discussions would have resulted from expanding and contracting columns in the mantle below without appreciable exclusion of magma. The eugeosyncline, on the other hand, has been built partly by volcanic activity. This is part of the unrest of the continental margin, and for some reason the mantle there has been, since Ordovician time at least, the site of excessive heat evolution causing magmatism and surficial orogeny.

Magmas of the Latite Ignimbrite Subprovince

Petrology. The first requirement in consideration of magmas of the latite-ignimbrite subprovince is a voluminous supply of a fairly uniform quartz latite magma. The volume is comparable to that of the Columbia River basalt field. The composition appears fairly uniform within the province; according to Howel Williams a number of rocks called andesites and dacites are only such by certain systems of nomenclature, and are really close to the quartz latite welded tuffs. Certain stocks are as basic as diorite or quartz diorite, and provision for them must be made in any theory of origin of the magmas.

Relation of Welded Tuffs to Stocks. The commonness of monzonite and quartz monzonite stocks and their similar composition to the quartz latites is striking. Nothing can be added to Gilluly's discussion of the close relation of the two as reviewed on previous pages, in which he postulated a reservoir of primary magma of quartz latite composition from which both the intrusives and extrusives were derived without further differentiation. Stringham's survey of the stocks of western Utah, Nevada, southern California, Nevada, and New Mexico indicates that some are as basic as diorite, but these are few. The quartz diorite of the Cottonwood stock of the central Wasatch (Fig. 38.2) lies close geographically to the Bingham quartz latite or granite stock and indicates the variation in composition that can exist within a few miles. As to the distribution, the stocks are abundant in the ignimbrite subprovince but equally abundant, evidently, outside the subprovince but within the Basin and Range province. It is computed that one stock occurs in about every 100 square miles, on the average, and each stock has an exposed area of about 5 square miles. From data at hand no difference in composition can be noted in the intrusive rocks inside the subprovince from those outside, but possibly there is a small difference which has not been detected.

Stringham (1958) has classed the stocks in two divisions, the aphanitic matrix porphyry and granitoid. The first he regards as mobile intrusions, but the second he believes formed by granitization. The Cottonwood stock of Fig. 38.2, for instance, formed by granitization, and the Bingham stock was probably intruded.

The age of the stocks is approximately the same as the welded tuffs. Some stocks are as old as Eocene, and others as young as Miocene according to zircon and potassium-argon age determinations, so they seem to predate, possibly accompany, and postdate the great avalanche eruptions. For instance, the Cottonwood (Alta) stock of the Wasatch Mountains is late Eocene (Crittenden *et al.*, 1952), the Sheeprock stock of the Sheeprock Range is middle Miocene or 15 to 17 m.y. (Cohenour, 1957), and

the Silver City stock of the East Tintic Mountains, mid-Eocene or 38 to 46.5 m.y. (Morris, 1957). No one has proposed that the stocks may have fed the welded tuff flows because the stocks have generally been considered older than the volcanics, yet the new age determinations indicate some may be younger. The welded tuffs are undoubtedly fissure eruptions.

It appears logical, therefore, to conclude that the ignimbrite magma was similar to that of the stocks, at least the porphyry stocks. Elsewhere in the general province a flow and a pyroclastic sequence of greater variability without the proponderance of welded tuffs and generally with a greater amount of basalt (olivine) occur.

Only one area within the ignimbrite subprovince of which the writer is aware has appreciable basalt. In a 6000-foot sequence of volcanics in one of the canyons of the Pioche, Nevada, district, is an olivine basalt unit 120 feet thick. The rest of the rocks are described as rhyolite, dacite and andesite, with the last two predominating. These are taken to be the welded tuff sequence, but some of them could be the younger Miocene-Pliocene volcanics.

Relation of Flows to Miogeosyncline. The latite avalanche subprovince is entirely within the miogeosyncline, with the exception of a very small overlap on the eugeosyncline southeast of Winnemucca. The western limit of the avalanche flows is close to the boundary between the eugeosyncline and miogeosyncline. The writer is not inclined to take this distributional relation to the geosynclinal divisions as very significant, because welded tuffs occur in the eugeosyncline of western Nevada and eastern California as part of the younger volcanics, and hence magma of avalanche composition and propensity can form in the crust under the eugeosynclinal strata.

Relation of Flows to Gravity Faults. The Wasatch and Cache Valley faults extend the Basin and Range system into the trench faults of southeastern Idaho and westernmost Wyoming from where they continue northward through Montana to the trenches of British Columbia with very little volcanism evident. The faults of the High Plateaus of Utah extend southward beyond the welded tuff subprovince.

Basin and Range faulting is believed to have started in about early Oligocene time, at least in southern Nevada, and to have continued from

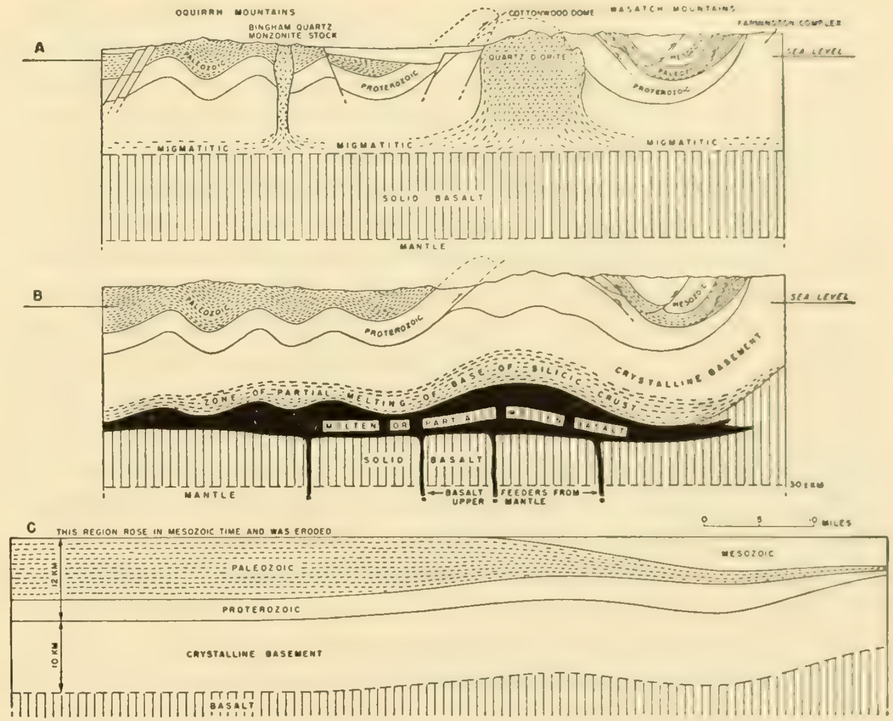


Fig. 38.2. Postulated origin of the monzonite-latite magmas. Section A is of the Oquirrh and Wasatch Mountains, Utah, and is factual at the surface but interpretive at depth. Section B represents the Laramide folding before the quartz monzonite and quartz diorite intrusions. Section C is section B restored to pre-folding time.

place to place to the present. Since early Oligocene is the time of the major avalanches a tie may be imagined. The flows are widely broken and tilted by the faults. After considerable erosion they were covered by the later volcanics and associated sediments, which in turn have been broken in places by further faulting. The association implies a genetic relation of the volcanics and gravity faults, but close scrutiny leaves one with the thought that the association is not as ubiquitous as desired.

Relation of Laramide Structures to Crustal Velocity Layers. In order to approach the problem of the origin of the latite magma, the relation of the Laramide structures to seismic velocity layers must be

considered. This has been done in Chapter 31 and Fig. 31.26.

If the boundary layers are fairly flat, as suspected from previous geological analysis, then widespread adjustment of the deeper crystalline basement must be postulated incident to the folding and thrusting of the Paleozoic and Mesozoic basin sediments above. The sections of Fig. 38.1 have been prepared to show the folding and faulting where seismic information on crustal layers is most available. Section B extends from the blast site at the south end of the Promontory Range westward to Elko, Nevada. Graduate theses at the University of Utah furnish stratigraphic and structural information on Promontory Range (Richard Olsen), Newfoundland Range (R. E. Paddock), Silver Islet Range (Fred Schaeffer and W. L. Anderson), and Pilot Range (Donald Blue). Sharp's (1942) mapping of the Ruby Range and Dott's (1955) work at Elko and eastward permit the drawing of a fairly satisfactory, if somewhat generalized and simplified, cross section. The 9-kilometer surface, if of uniform depth across the entire section, is just about tangent to the troughs of maximum downfolding of the Paleozoic and Proterozoic sedimentary sequences. A more detailed and larger-scaled section is shown in Fig. 38.2 which takes a course northeasterly across the Oquirrh Mountains, across the Jordan (Salt Lake) Valley to the Cottonwood dome of the Wasatch Mountains, and then northward to the major exposure of the Archean crystalline (Farmington) complex. If the transition surface between the silicic and the basaltic crust is as illustrated in restored sections B and C, then considerable flowage must have taken place in the base of the silicic crust during and after folding of the sediments.

Origin of Latite Magma. The chief reason for postulating the primary nature of the latite assemblage of the Great Basin is great volume with only minor amounts of rocks of other composition. A few basalt flows have been noted as part of the latite assemblage but most basalt occurrences are later, and were extruded in Pliocene–Pleistocene time. Andesite, dacite, and rhyolite generally occur along with the latite, but in relatively small amounts. A plausible theory for the origin of the magma must therefore account for variations as indicated by the above observations, and even, on occasion, to explain the transit of basalt to the surface.

Two possibilities occur to the writer: (1) the base of the silicic crust melts in part or in bulk to form the primary magma, or (2) basaltic magma is intruded in megasills at the base of the silicic crust at temperatures sufficiently above the melting temperature of the silicic crustal rock to melt an appreciable layer of it or to melt it partially in decreasing amounts upward from the basalt sills; some mixing of the basalt with the melted silicic crust might occur. The second theory supplies heat for the phenomenon and basalt, on occasion, as required. An expanding column of the mantle to produce a surface uplift of about 2 kilometers is needed everywhere in the Rocky Mountains, Colorado Plateau, and Great Basin region (see Chapter 31), and a primary basalt magma is needed under most of it, so it seems logical to start with the premise of rising basalt from the mantle where, it has been concluded, differences in density exist. The basalt would furnish a good part of the heat needed to raise the temperature of the base of the silicic crust to melting. The idea of *partial* melting of the base of the silicic crust, especially those parts thrust slightly downward during the Laramide orogeny, is attractive because, thereby, a magma of monzonitic composition might be formed rather than one of granodioritic composition as in the case of bulk melting of great roots. Partial melting will not only facilitate viscous flow to level out the base of the silicic crust (Fig. 38.2) but also will produce the great volumes of various gneisses and schists called migmatites whose features characterize them as transitional to igneous. A granitic or monzonitic magma would have been squeezed out, and represent the first minerals to melt, hence more acidic, and more basic varieties would represent the melting of a larger percentage of a basal portion of the silicic crust nearby. The basalt immediately beneath may be tapped by a fissure conduit from time to time and add its conspicuously dark and perhaps unexpected presence to the surface assemblage.

Mixing of a small amount of the silicic magma with basalt would produce an andesite, or the basalt could fractionate to an andesite. Very little andesite is needed in this province.

The latitic magma of the ignimbrite subprovince contained sufficient water such that effervescence of water vapor at a temperature high enough for welding occurred. The extrusion temperatures of the tuff-breccias in the Pine Valley Mountains is interpreted to be lower than that necessary

for welding, and perhaps the ignimbrite subprovince was determined not only by abundant water but also by a higher than normal temperature. It corresponds to the postulated projection of the East Pacific Rise under the western United States (Chapter 31).

TECTONO-IGNEOUS PROVINCES AND DEEP-SEATED EARTHQUAKES

The South American Andes and adjacent shields and basins are noted for their intermediate depth and deep-seated earthquakes. In charting the foci Benioff (1954) was led to the conclusion that they lie along an extensive, inclined plane or surface that extends down under the Cordillera and stable region to depths of nearly 700 kilometers. He illustrated the earthquake foci along two sections, one across northern Chile and Argentina, and another across Equador to the Guayana shield. For these sections the writer has idealized the crustal geology as shown in Fig. 38.3 in accord with the more detailed cross sections of Fig. 34.5. It will be seen that the volcanic fields as previously pointed out are east of the Nevadan batholithic belt and lie principally on the deformed miogeosyncline. They occur somewhat shoreward of the break in slope of the earthquake foci surface. Benioff postulates this surface to be a gigantic reverse fault due to compression in the mantle. It has also been postulated that this great fault is the region of origin of basaltic magma, especially of the alkali olivine variety (Kuno, 1959). Here is a likely place where the partial fusion of the upper mantle shell occurs to supply basalt and heat to the subcrust and crust, and where consequent intrusive and extrusive magmatic activity takes place. Although andesites are widely recognized in the Andes, Dr. Howel Williams informs the writer that he has a knowledge in part and a strong hunch that great volumes of the volcanic piles in South America and Mexico are of the composition called latite or quartz latite in the discussion of the Great Basin volcanics on previous pages. If the base of the silicic crust is fused partially, then by postulate, more latite than andesite would probably be extruded.

The deep-seated earthquakes in South America and Mexico are complemented by a trench at the continental margin, which is presumed by Benioff and others to be a compressional consequence of reverse movement along the great fault defined by the earthquakes. Deep-seated earthquakes have not been recorded in the western United States or western Canada, and no trench exists at the continental margin; yet, the other igneous and tectonic components of the western orogenic belts are present. It seems to the writer that the deep-seated earthquakes have been an integral part of the western Cordillera of Canada and the United States during most of the Tertiary, but that the fault along which they occurred is now inactive. It may have been replaced by the East Pacific Rise and associated expansion of the mantle.

CRUSTAL TENSION AND MAGMATISM

Previous References

The belief that the earth's crust has suffered large amounts of shortening in the orogenic belts has been an orthodox tenet of geologists for many years. Lately several individuals, including de Sitter (1956) and Bucher (1956), have argued for vertical uplift with consequent gravity flow or sliding of the surficial rocks away from the uplift to form the folds and thrusts, and therefore, for minor amounts of, or no horizontal shortening. Others are now contending for expansion of the earth and tension as the primary and dominant force of crustal deformation.

In Chapters 41 to 43 the concept that the earth is expanding is mentioned in connection with the possible drifting apart of North and South America. Also in Chapter 31 the Basin and Range province was explored relative to tension in the crust and expansion of the earth. The Mid-Atlantic rift in Chapter 10 was treated as a tensional structure as a result of earth expansion. It is now absorbing to speculate on magmatism in the framework of crustal tension.

Evidence of Tension

Fissure Eruptions and Tension. The most plausible concept of the origin of fissures through which large volumes of magma have passed to the surface is one of tension. Fissure eruptions have always been difficult to explain in the framework of crustal compression. Fissures through which basic magmas in large amounts have flowed from the basaltic subcrust

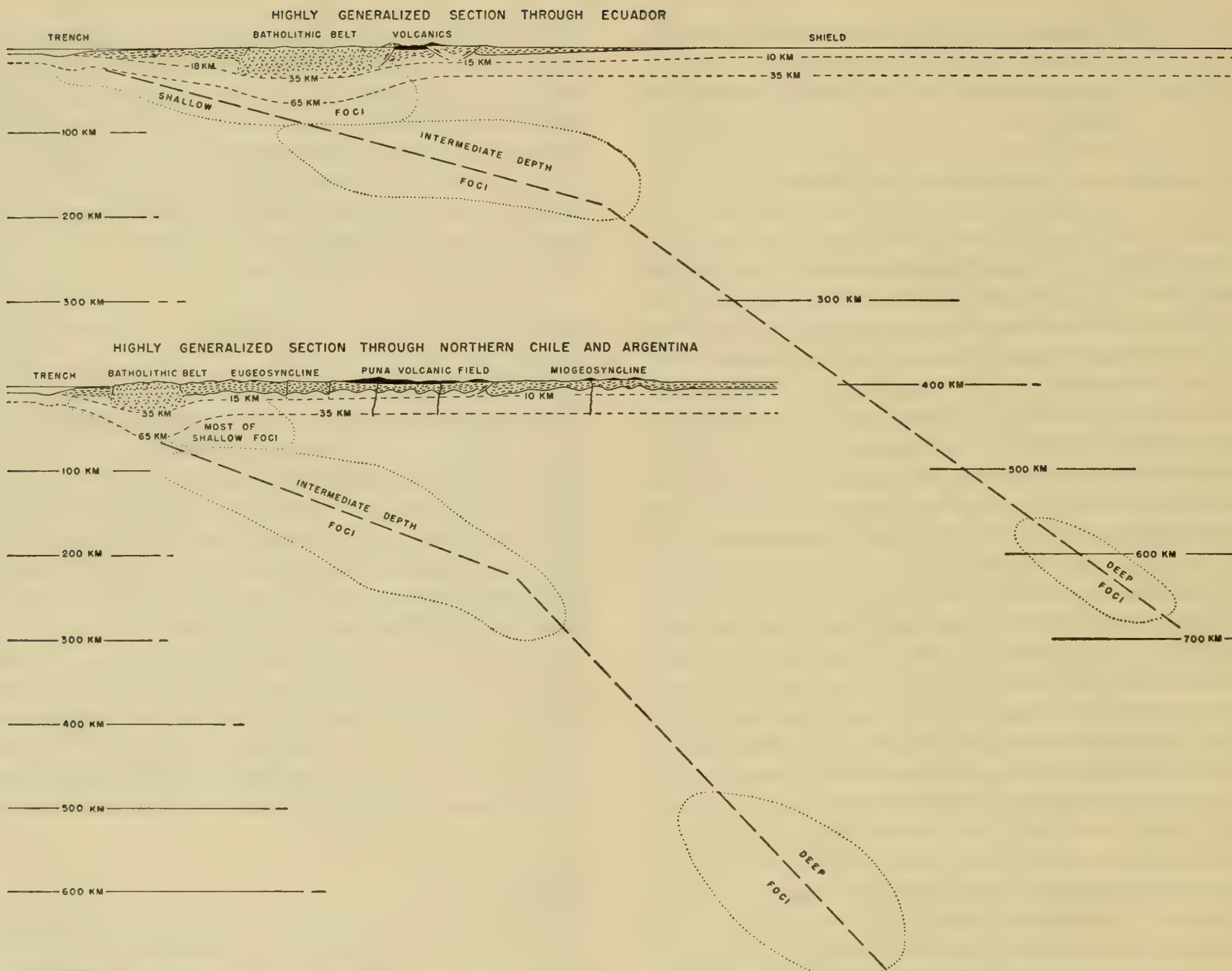


Fig. 38.3. Relation of deep-seated earthquakes, postulated fault in mantle, and crustal constitution in the South American Andes.

to the surface cannot be explained as surficial features of the folding of sedimentary sequences.

Basin and Range Province and Tension. In Chapter 31 it was postulated that the Basin and Range province has been distended about 30 miles since Miocene time, and the suggestion made that this process could provide for the intrusion from great depth of much magma. The earthquake foci have been interpreted to mean that the great faults extend to depths of 20–40 kilometers.

Eugeosyncline and the Tension Hypothesis

With an expanding earth and a crust cracking apart in places we may devise a scheme of magmatism for the eugeosyncline. See Fig. 38.4A. The crystalline crust of the continents seems to end abruptly at the ocean basins, and in the realm of an expanding earth the continent-ocean boundary may generally be a zone of weakness where extension will be focused. If so, then here will be a likely site for the rise of magma from the upper mantle. The continental margins are commonly sites of deep-seated seismicity as well as unusual thermal activity in the mantle. Fissure eruptions in the eugeosyncline have been postulated.

Provision must be made for the evolution of the andesites, and if they arise by fractional crystallization from basalt, then there must exist large magma chambers in the subcrust where the process takes place. It would appear that basaltic eruptions should be dominant in the early stages of the eugeosyncline with andesites more abundant later; also undifferentiated basalt could be conducted to the surface from time to time as new fissures break through to great depths. It is not known if observations in the eugeosynclines support the supposition that andesites become more abundant in the later stages.

Batholithic Belts and the Tension Hypothesis

Speculating further, as the eugeosyncline develops the crust is depressed under it, and the depression is mostly the result of removal of support by the ejection of magma through fissures to the surface. See Fig. 38.4B and C. Eventually, the silicic upper crust or the base of the eugeosyncline comes into the domain of melting, and it is at this stage, with continued

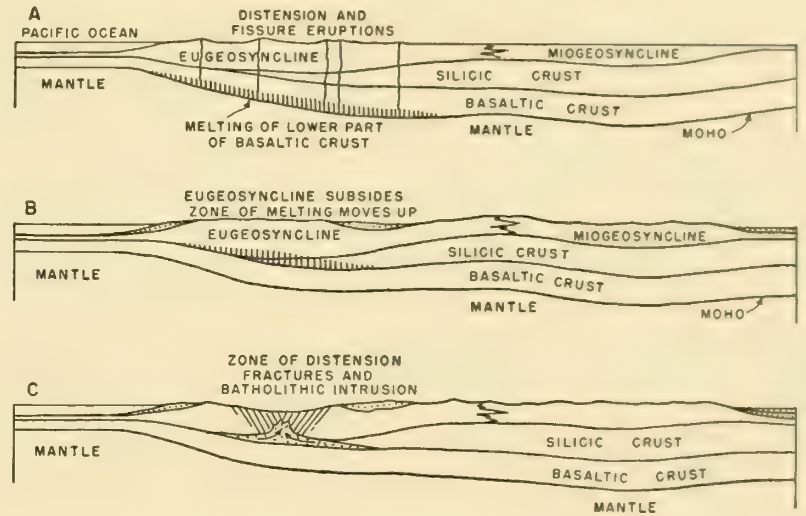


Fig. 38.4. Speculations on crustal structure at the continental margin and the relation of magmatism to the eugeosyncline and batholithic belt, if the earth should be expanding and the crust distended.

expansion and tension that the growth of the batholiths of intermediate and acidic composition begins. The fluidity of the basaltic melts provides for rapid flow to the surface, but the greater viscosity of the more silicic magmas makes for slower, more irregular intrusions, with attendant varied intrusive relations. The space problem is largely accounted for, however, by irregular fissuring and pulling apart of the crust (Fig. 38.4C).

Several adjacent fissures may develop, and each is invaded by the silicic magmas, thus perhaps accounting for the great septa of metamorphosed country rock noted in some of the batholithic belts. The batholithic belts in some places are narrow and linear and seem to fit nicely the tension hypothesis, but others are more irregular with the batholiths in clumps, and therefore do not accord with the hypothesis very well.

Problems

The conventional explanation for the origin of large volumes of magma of intermediate composition is the melting of the lower part of a thick-

ened silicic crust; the thickening is due to compression. In the concept of expansion and tension no thickening is possible. In fact, the conditions of rise of the batholithic magmas may be similar to those of the quartz latite magmas of the Basin and Range province where the lower part of the silicic layer is believed to melt without previous thickening.

The explanation of batholithic belts based on extension fails to provide adequately for dynamic metamorphism and isoclinal folding such as occurs in the Sierra Nevada. Possibly a facility for such metamorphism is present in the slices of rock that settle toward the batholith as extension occurs. See Fig. 38.4. It would appear that an extensive aureole of thermal or hydrothermal alteration would occur around the intruding batholith because of the fracturing. In the Andes of Peru no stage of dynamic metamorphism is reported just prior to batholithic intrusion (H. L. Hosmer, personal comunication), so the problem evidently does not exist there.

Another phenomenon that occurs and for which an explanation is not readily seen by the writer is the post-batholithic uplift. The belts of batholithic intrusion are elevated and deeply eroded to expose the large intrusive bodies. In the framework of crustal extension what causes the uplift? One may counter that the Sierra Nevada block is a much later affair and not related to the uplift immediately after intrusion. Also it may be observed that parts of the batholithic belt of South America and North America are fairly low-lying today and that the batholiths may have come closer to the surface than illustrated in Fig. 38.4.

Still another problem is apparent in consideration of the Gulf of California. It was speculated that the eugeosyncline forms by fissure eruptions as the crust is pulled apart. In Chapters 30 and 31 evidence was presented that Baja California has been pulled apart from the mainland of Mexico during Cenozoic time, so we should expect extensive fissure eruptions there. The volcanism instead is concentrated on the east in the Sierra Madre Occidental with some also on the west in Baja California. Only a few volcanic cones exist in the Gulf itself. Possibly no large magma chambers existed where the fractures and separation occurred.

Basic Conflicts

The foregoing discussion of the origin of the various magmas is wrought with several conflicting concepts. In certain considerations we conjure up a state of compression in the crust and outer mantle; in others we entertain extension of the crust. Where extension, we recognize certain zones of extension complementary to zones of compression, or we imagine world-wide tensional strain. If tension in local zones, then we usually think of convection circulation in the mantle; if world-wide, we say expansion of the entire earth. Some believe a little expansion has occurred, some considerable, but considerable expansion appears impossible (Cook and Eardley, 1961). Others recognize local or regional vertical uplift due to changes of state in the mantle as the basic tectonic activity, without appreciable overall earth expansion. Secondary, gravity-caused flow movements on the flanks create the compressional structures. World-encircling rises underlain by an expanded mantle-crust transition layer seem to be a reality. And finally, there are many geologists who support the concept of drifting and rotating continents without earth expansion. These movements are commonly attended by horizontal coupling of varying magnitudes. Then there is the pointed conflict of granitization versus magmatic intrusion, particularly in regard to the great batholiths of the eugeosyncline. The writer finds convincing examples of each and all of the above-mentioned theories, yet none seems adequate to explain the entire panorama of structural and igneous observations.

It was hoped that the igneous rocks, when their origin was investigated and related to crustal structure, would point out which of the theories are valid, and perhaps the study has accomplished this to some small extent, but there still remains much uncertainty.

39.

ALASKA AND THE YUKON

GEOMORPHIC PROVINCES OF ALASKA

The principal geomorphic provinces of Alaska are, from north to south, the Arctic Coastal Plain, the Brooks Range, the Central Yukon Plateau and Lowland, the Alaska and associated Coast Ranges, the Alaska Peninsula, the Aleutian Archipelago, and the Alexander Archipelago. See map, Fig. 39.1. They are part and parcel of the continent's great western Cordillera. The generalized tectonic divisions are shown in Fig. 39.2.

The Brooks Range stretches east-west across northern Alaska and includes several smaller ranges, such as the De Long, Baird, and Endicott Mountains. They support an extensive upland erosion surface, whose higher elevations reach from 5000 to 6000 feet above sea level. The Brooks Range in its central portion is a sharply defined mountain mass that rises conspicuously from the Yukon Plateau on the south and above the foothills of the Coastal Plain on the north. The Colville River drains much of the northern slopes of the range and the piedmont of the Arctic Coastal Plain. The Brooks Range is covered for the most part with perennial snow fields and contains a number of glaciers. The erosional features are described as distinctly youthful, and presumably very little erosion has occurred there since the once far greater ice fields and valley glaciers of the Pleistocene have disappeared. The Arctic Coastal Plain from the air appears as a bleak, flat wasteland of frozen lakes and rivers and snow-covered flats.

The Yukon or Central Plateau in the central and upper Yukon drainage is a broad dissected plateau bounded on the north by the Brooks Range and on the south by the Alaska and Coast Ranges. The two great ranges are about 300 miles apart. The plateau loses definition in the lower Yukon drainage, where it is characterized by the flat-topped interstream areas separated by broad and low-lying, estuarine-like embayments. A few minor ranges and peaks rise above the general level of the upland surface. The Yukon River has eroded a meanderbelt 35 miles wide in places, but with several narrows along its coarse. Near its mouth, a very low alluviated portage separates it from the Kuskokwin River, and both rivers are in the process of building large deltas in the Bering Sea.

The Central Plateau was dissected and then, during the maximum glaciation, heavily alluviated, chiefly with silt. The Yukon and tributaries have since been engaged mostly in clearing out the silt.

The Seward Peninsula is a geological entity in itself and will receive special mention later. It is generally included in the Central Plateau province.

The Coast Range of southeastern Alaska and British Columbia extends northwestward by way of the St. Elias Range, Wrangell Mountains, and Nutzotin Range into the Alaska Range, which together form a great arc approximately parallel to the margin of the Gulf of Alaska. The Alaska Range continues southwestward to the Aleutian Range, which forms the backbone of the Alaska peninsula. Mt. McKinley in the Alaska Range (20,300 feet) is the highest mountain in Alaska. The St. Elias Range and the Chugach Mountains support the greatest ice field in North America; several peaks rise above 14,000 feet, including Mt. Logan, the highest at

Fig. 39.1. Index map of Alaska and the Yukon.

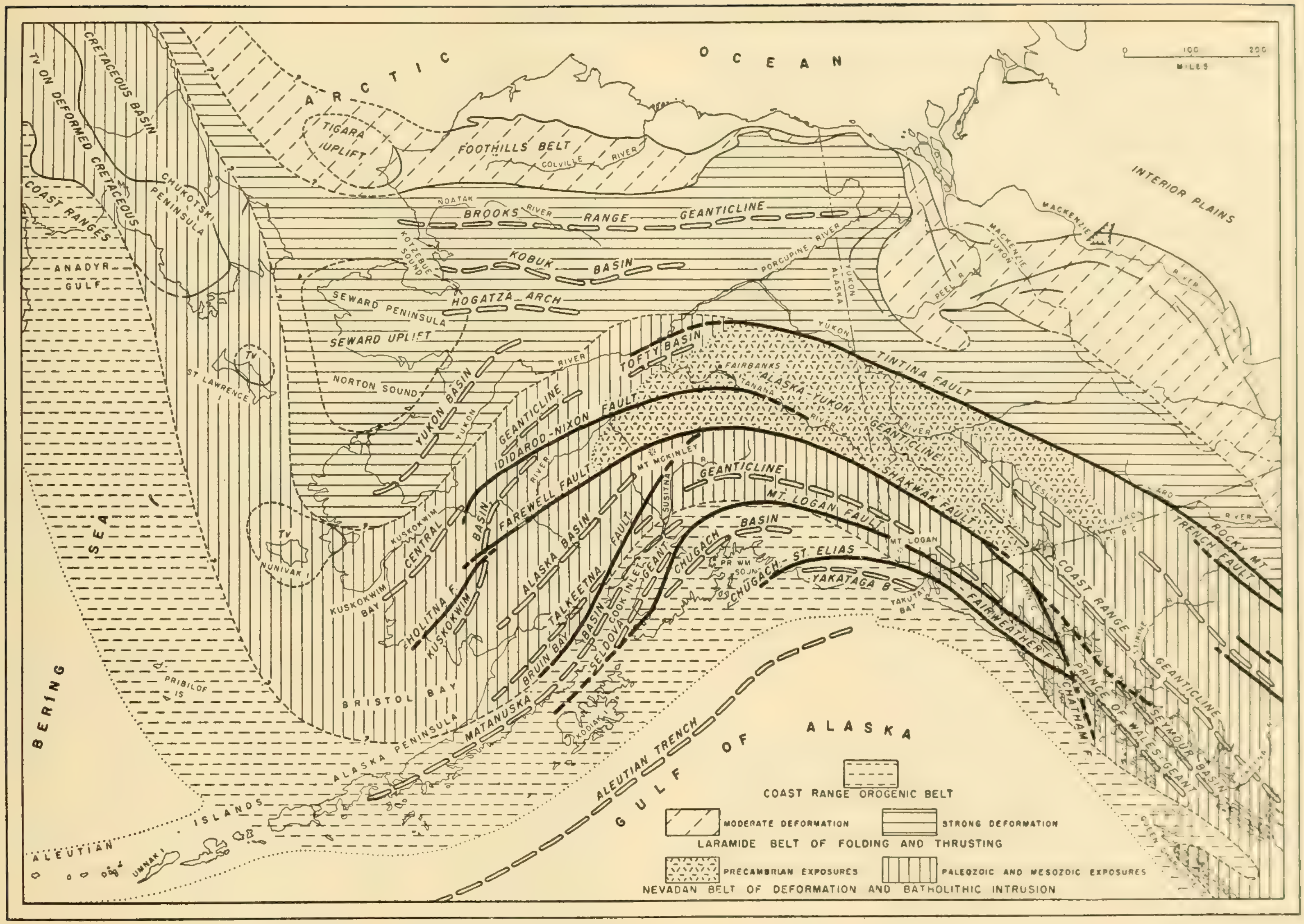


Fig. 39.2. Tectonic map of Alaska and the Yukon Territory. Laramide orogenic belt characterized by folds and thrust faults mostly in shelf and miogeosyncline sedimentary rocks. The structures of the basins are Late Cretaceous or Early Tertiary; of the geanticlines are partly of Early Cretaceous age or older, Nevadan orogenic belt characterized by numerous batholiths and deformed eugeosynclinal sediments. The intrusions and structures are Early and Late Cretaceous with Early Tertiary structures in the Alaska Range area. Coast Range orogenic belt is marked in part by Tertiary sediments and by folding and thrusting principally during the Late Cenozoic. Interior Tertiary basins not shown.

19,850 feet. Mt. Sanford in the Wrangell Mountains is 16,000 feet high. The Aleutian Range has a general summit level of 3000 to 6000 feet and includes a number of active and dormant volcanoes. The famous Valley of Ten Thousand Smokes and the great crater of Aniakchak are situated at the southwest end of the Aleutian Range.

An oceanward arc of ranges, more truly called coast ranges, extends through the Cook Inlet and Prince William Sound region. The Kenai Mountains forming the Kenai peninsula east of Cook Inlet connect with the Chugack Range, bordering the coast, and this merges with the St. Elias Range. Some writers have grouped the St. Elias Range entirely with the Coast Ranges, but its geology is too little known to permit a definite conclusion. At the west end, the Kenai Range probably is continued by the low mountains of Kodiak Island, and thence by beveled or buried elements in the shelf between the Aleutian Islands and Aleutian trench to Unimak Island.

The Shelikof Strait–Cook Inlet and Susitna River depression effectively separates portions of the inner ranges from the outer. The Copper River Valley and its tributary, the Chitina River, form another separating depression. The Talkeetna Mountains break the continuity of the two depressions and, anomalously, seem to bridge the two great mountain systems.

PALEOZOIC GEOSYNCLINE AND RELATED OROGENY

Most of the Paleozoic rocks of Alaska are exposed in the Brooks Range, Seward peninsula, Central Plateau, and Alexander peninsula. The latter has already been considered in a previous chapter. The Alaska, Nutzotin, and Wrangell Ranges also contain Paleozoic rocks, and a nearby belt extends along part of Copper River and Chitina River valleys. The towering mass of Mt. McKinley in the Alaska Range is eroded mostly from deformed Paleozoic strata.

For a detailed study of the Paleozoic rocks of Alaska, Smith's *U.S. Geological Survey Professional Paper 192* should be consulted, particularly the large correlation chart in the pocket. The formations are well developed in the Tanana-Yukon region of the Central Plateau, and a résumé of them is given on page 609.

The igneous rocks of the Tanana-Yukon region have been summarized by Mertie (1935). Basic lavas of basaltic and diabasic character have been extruded during at least five geologic epochs in the Paleozoic. The first was in the Middle Ordovician, the second in the Middle Devonian, and the last three during three epochs of the Carboniferous. Granular intrusives of the same general character accompanied the extrusion of the lavas, but the volume of such rocks is relatively small. Some rhyolite and dacitic lavas and tuffs are found among the Carboniferous lavas, but generally speaking, lavas of acidic or intermediate character are rare. Ultrabasic rocks were intruded during the late Devonian epoch.

The volcanism that occurred during the Carboniferous period in Alaska, according to Mertie (1935) was greater than in any other period and most intense in the Alaska Range. The eruption of the basic lavas was accompanied by epeirogenic movements that persisted into the Triassic.

It is immediately clear that the above rocks represent the eugeosynclinal assemblage previously recognized and described in the western Cordillera of southeastern Alaska, British Columbia, Washington, Oregon, California, and Nevada. The presence of the basic intrusives in the volcanic assemblage suggests that the belt was more the site of the archipelago than an adjacent trough.

In northern Alaska, the Paleozoic rocks are mainly sandstones, shales, and limestones, and are typical of the miogeosyncline or shelf, also previously described in the Cordillera of Canada and the United States. No volcanic rocks have been found in the sediments. A résumé, as listed in the correlation chart of *Professional Paper 192*, is given on page 610.

The Upper Devonian and Mississippian rocks of a southeastern area of the Brooks Range have recently been measured, and the section is given in Fig. 39.3. Although about 8000 feet of strata of the two systems are present, they are regarded as platform-type deposits and not miogeosynclinal by Bowsher and Dutro (1957). The massive lower and middle members of the Kanayut conglomerate help to define a region of uplift in the Late Devonian. See Fig. 39.12. The above strata are overlain by variegated shale and siltstone, the Siksikpuk formation, about 350 feet thick, which is probably Permian in age, and then the Shublik formation of Triassic age. The Pennsylvanian is missing over all Alaska except the northeast corner.

Selected Paleozoic Sections of Alaska		Thickness, Feet
ermian		
Kandik district	Limestone; Conglomeratae, shale, and sandstone — Tahkandit limestone	
ississippian		
Porcupine district	Dark shale and limestone, in part same as Calico Bluff formation	?
Koyukuk-Melozi district	Greenstone, little rhyolite, formerly considered part of Kanuti group	?
Yukon-Tanana district	Clay shale, sandstone, conglomerate. Nation River formation	4–6000
	Lava flows and associated sediments. Rampart group and Circle volcanics	5–10,000
	Limestone, shale, slate. Calico Bluff formation	13,000
	Limestone beds	
	Undifferentiated schist, shale, chert, quartzite	
	Chert with minor amount of limestone and shale, Livengood chert	2–4000
Wiseman-Chandalar district	Comparable with northern Alaska section	
Marshall district	Andesite and basalt flows and tuff	?
Sheenjek district	Dark limestone, somewhat silicified, weathering light, with argillaceous and arenaceous beds	6000 plus or minus
	Quartzite, conglomerate, shale, with cherty matrix	
Chisana district	Clay slate; Shale and conglomerate; Conglomerate — Wellesley formation	1–2000
evonian		
	Upper	
Wiseman-Alatna district	Quartzite, sandstone, slate, little conglomerate, grit, limestone	?
Wiseman-Chandalar district	Quartzite, flint, calcareous black slate, impure limestone. Formerly part of West Fork formation	?
Eagle district	Basalt lava and pyroclastics of greenstone habit. Woodchopper volcanics	10,000 plus or minus
	Middle	
Porcupine district	Brown shale and basalt	
	Massive light gray-blue limestone. Salmontrout limestone	300
Kandik district	Argillite, chert, and cherty grit	

Selected Paleozoic Sections of Alaska		Thickness, Feet
	Thin beds dark gray limestone, shale, and chert	1000
	Lithographic limestone, dark gray crystalline limestone	500
Central-East district	Clay shale, siliceous slate, chert, quartzite, sporadic limestone, and conglomerate	?
Wiseman-Chandalar district	Slate with small amounts of limestone	
Kantishna-Nenana district	Massive limestone, equivalent to part of Tonzona	10,000 plus or minus
	Limy shale, more calcareous at top	
	Slate, argillite, graywacke, quartzite	
	Black conglomerate, white conglomerate, shale and graywacke	
Chisana-Tok district	Crystalline limestone associated with black slate, sandy beds, somewhat schistose	?
Sheenjek district	Quartzite, sandstone, slaty sandstone, and argillaceous sediments	?
Silurian		
	Middle	
Porcupine district	Black fissile shale, little siliceous limestone	
	Buff magnesian limestone	2500 plus
Sheenjek-Alatna	Slate, schist, thin layered limestone; Skagit	?
Kandik district	Massive somewhat siliceous limestone; Skagit	?
	Massive white to cream limestone	?
Preacher-Tolovana-Hot Springs district	Calcareous and dolomitic limestone, somewhat recrystallized. Tolovana limestone	3000 plus
	Lower	
Ruby district	Undifferentiated limestone	?
Ordovician		
	Middle	
Porcupine district	Gray limestone	600
Ruby district	Magnesian limestone overlain by calcareous limestone, not differentiated on map	5000
Preacher district	Volcanic tuff and associated igneous rocks	?
	Black shale, merging downwards into schist	?
Cambrian		
	Upper	
Kandik district	Limestone, with dark gray to black slate and chert in higher part	?
	Limestone	?
	Middle	
Kandik district	Upper plate of limestone	
	Thin layers of slate an quartzite	?
	Lower plate of limestone	

		Thickness, Feet
Mississippian		
Central-Western district	Massive light-colored semicrystalline limestone, considerably silicified; Lisburne limestone	4000 plus
	Sandstone, shale, thin limestone, in place chert, conglomerate; Noatak formation	thousands
Canning district	Gray and black limestone, somewhat brecciated, much silicified, equivalent of part of Lisburne limestone	3000
	Black limestone, slate, shale and sandstone; age uncertain; rests unconformably on highly metamorphic schists	?
Devonian		
	Upper	
Colville-Noatak district	Quartzite, sandstone, slate, subordinate conglomerate, grit, and limestone	?
	Middle	
Lisburne district	Calcareous sandstones and shale	?
Canning district	Black shale, slate, and subordinate sandstone; not differentiated on map	1000 plus
Silurian		
	Middle	
Noatak-Kobuk district	Massive, somewhat siliceous limestone; Skajit limestone	6000 plus

A line separating the eugeosynclinal assemblage of sediments on the south from the platform (shelf) or miogeosynclinal sediments on the north follows the Yukon River approximately.

The distribution in outcrop and the geosynclinal thicknesses where known of the Paleozoic strata indicate that the whole of Alaska was a region of subsidence and sedimentation in the Paleozoic, and an extension of the Cordilleran geosyncline.

An episode of granitic intrusion occurred during the early Devonian in the North Fork of the Chandalar River along the south flank of the Brooks Range (Mertie, 1935). This is the only Paleozoic granitic intrusive so far identified in Alaska, and it is in the area of the mainland assemblage of stratified rocks. Representatives of the Lower Devonian are absent, and the strata of Middle Devonian age rest unconformably on those of Silurian age. The rocks below are more metamorphosed than those above the unconformity. This evidence indicates crustal unrest in the geosyncline, such as is found in Paleozoic beds of the Alexander

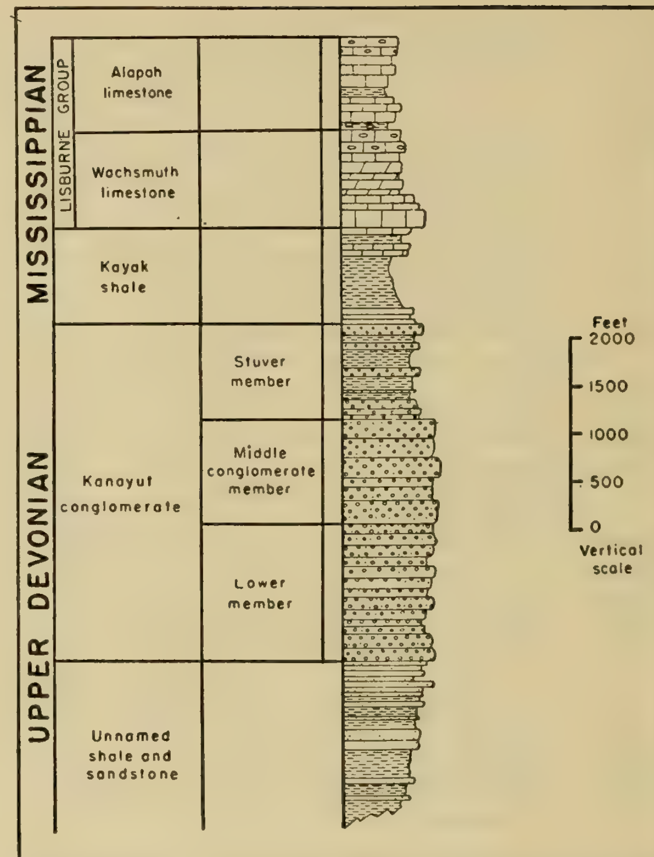


Fig. 39.3. Generalized section of Upper Paleozoic rocks in the Shainin Lake area. Reproduced from Bowsher and Dutro, 1957.

Archipelago, although slightly more continentward than the trough of accumulation of the volcanic assemblage.

Numerous other unconformities undoubtedly exist in the volcanic assemblage.

TRIASSIC AND JURASSIC GEANTICLINE AND ADJACENT BASINS

During the Triassic period and persisting into the Jurassic, a great geanticline rose from the Paleozoic geosyncline and separated two adjacent

basins of accumulation, one on the north and one on the south. Examine Fig. 39.4. The basin on the south collected chiefly sediments of the eugeosynclinal assemblage, while the trough on the north received limestones, sandstones, shales, and cherts of the miogeosyncline and shelf. Fixed lines have not been drawn on maps to show this feature, but Mertie (1930) describes it as a region of epeirogenic uplift and erosion.

The geanticline is a parallel, except in detail, of the Cordilleran geanticline in Canada and the United States, already described and pictured in the paleotectonic maps of Plates 9 to 11.

The columnal sections of Fig. 39.5 are characteristic of the Coast Ranges. Here in southern Alaksa Mid-Jurassic time marked the beginning of development of basins and separating geanticlines that persisted through the Cretaceous.

In the Kuskokwin region Cady *et al.* (1955) described the Gemuk group of siltstone and chert with local developments of basalt and andesitic rocks to be 15,000 to 25,000 feet thick.

The Triassic and Jurassic of the northern Brooks Range and Arctic Coastal Plain consists of two formations. The Shublik of Late Triassic age is 300 to 1000 feet thick and is composed of interbedded dense bituminous limestone, chert, shale, siltstone, limonite oolite, and calcareous glauconitic siltstone. It is entirely marine. The Kingak shale spans most of the Jurassic and is about 4500 feet thick. It contains graywacke, varicolored bedded chert, lenses of conglomerate, and coquina limestone.

CRETACEOUS BASINS AND GEANTICLINES

An examination of the geologic map of Alaska and the correlation chart of *Professional Paper 192* shows the Lower and Upper Cretaceous strata

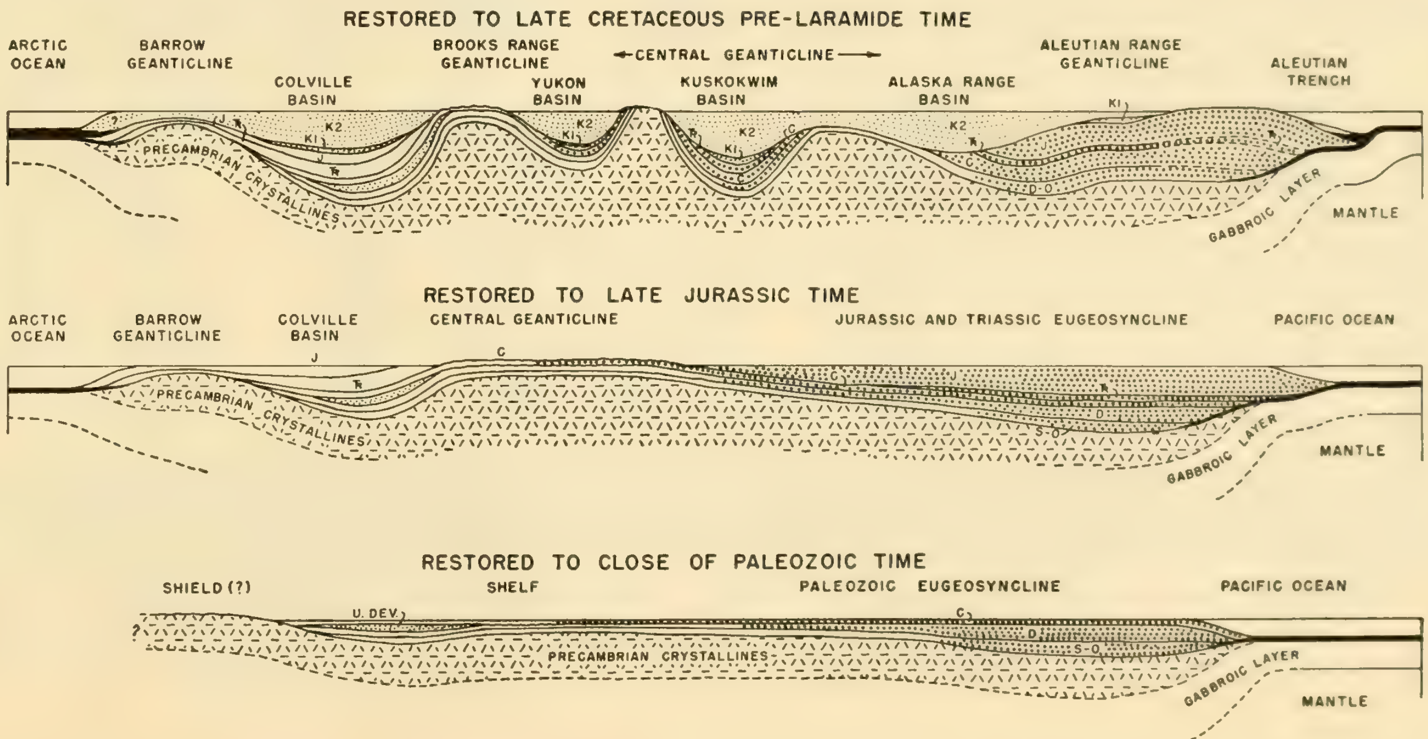


Fig. 39.4. Idealized evolution in cross section of Alaska from Point Barrow on the north to Kodiak Island on the south. In part after Cady *et al.*, 1955. Vertical scale highly exaggerated. Large dots indicate eugeosynclinal assemblage. Blank units represent miogeosyncline or shelf assemblage. Small dots represent clastics with dominant graywacke content. K1, Lower Cretaceous; K2 Upper Cretaceous. Nevadan structures and batholiths not shown. Mogatza arch and Kobuk basin between Yukon basin and Brooks geanticline not shown.

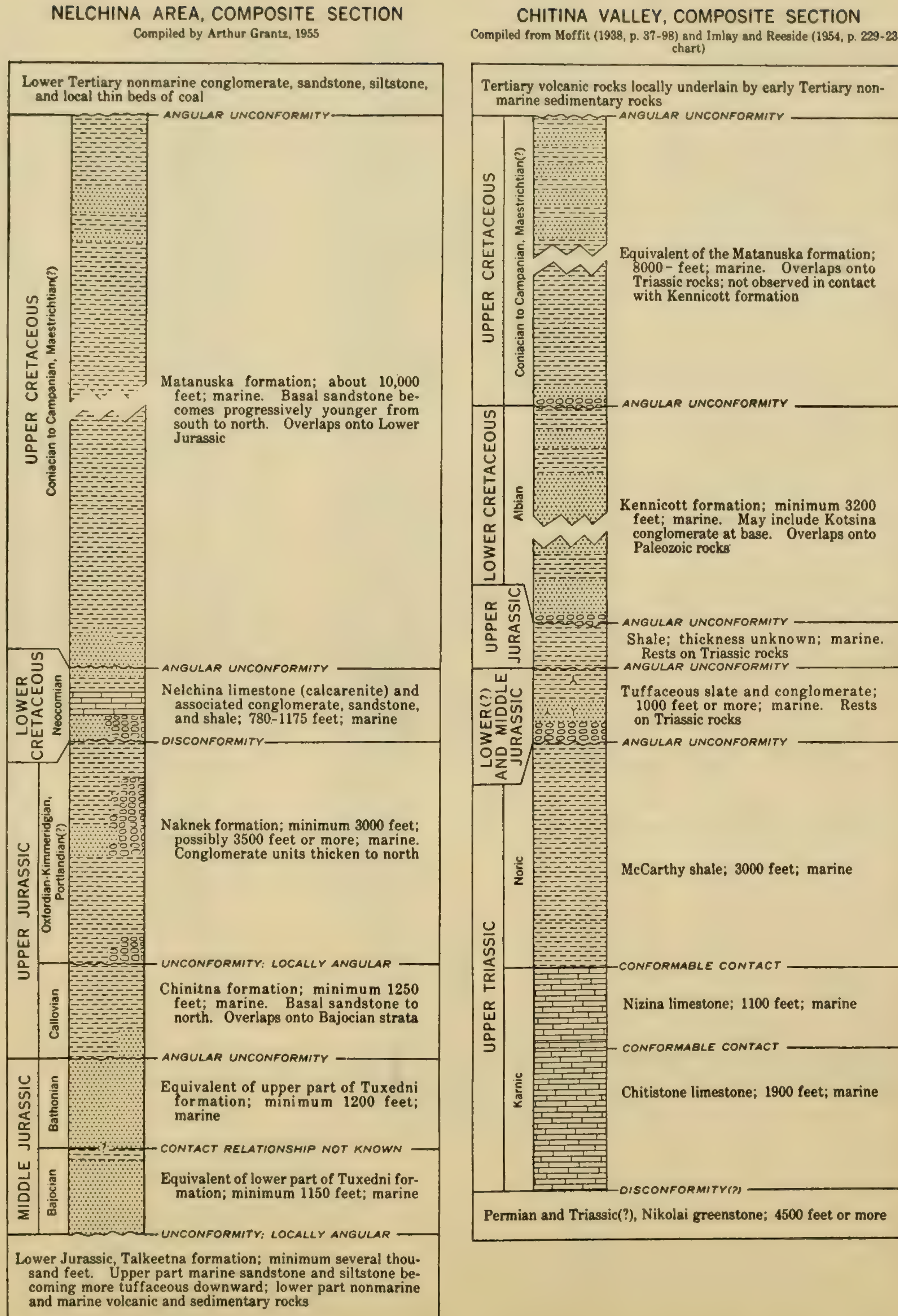


Fig. 39.5. Representative sections of the Cook Inlet Mesozoic province, southern Alaska. From Miller *et al.*, 1959.

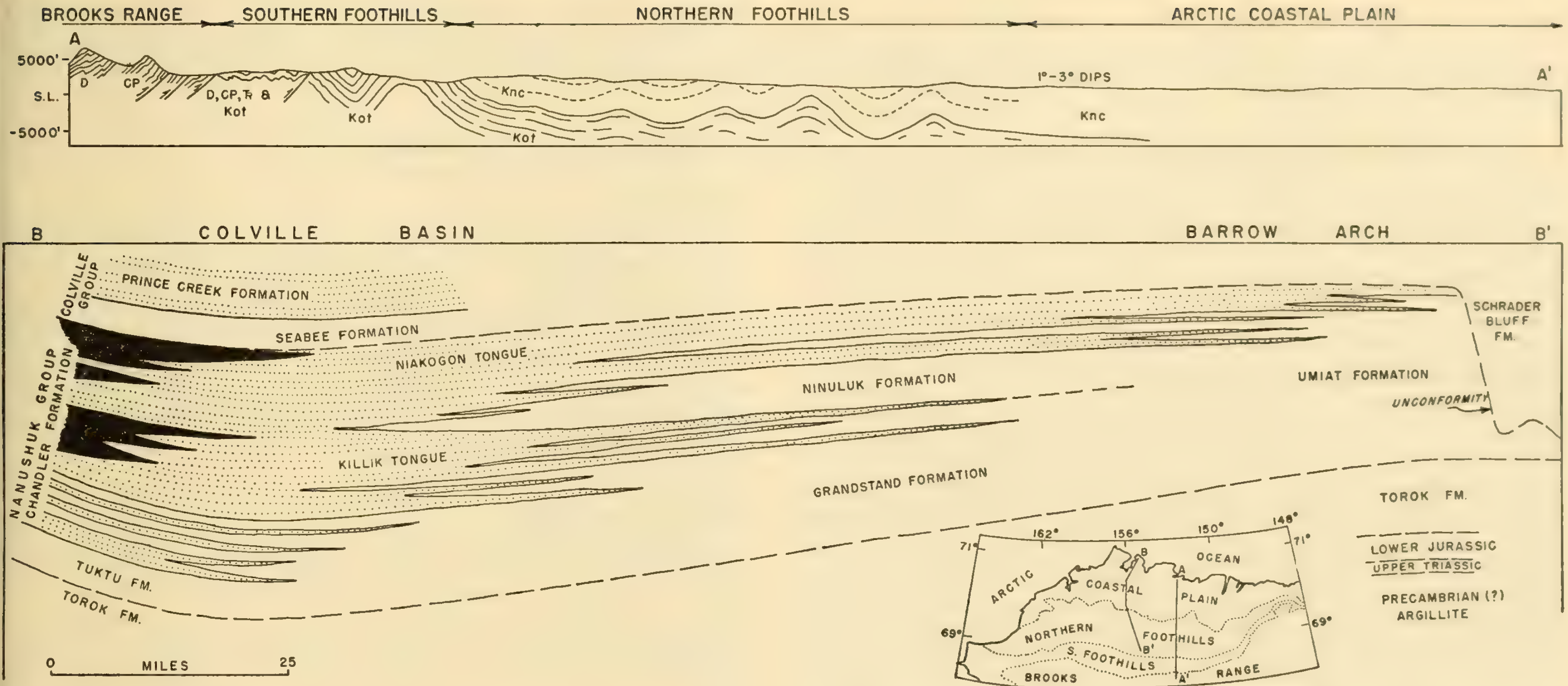


Fig. 39.6. Cross sections of Arctic Foothills, and Coastal Plain. A–A′, existing section; B–B′, restored to close Cretaceous showing facies of Nunushunk and Colville groups, after Payne *et al.*, 1951; somewhat altered after Grye *et al.*, 1956. Black is "inland facies"; stippled is "coastal facies"; blank is "offshore facies." Kot, Torok fm.; Knc, Nanushuk group.

so widespread that much of Alaska must have been under water and receiving sediments during Cretaceous times. Certainly large parts of the Triassic and Jurassic geanticline were covered. Mertie (1930) states that, at least at one time or another during the Cretaceous, all Alaska was subjected to sedimentation. But, it seems evident that a number of long linear uplifts rose and separated the basins of sedimentation in the manner shown on Fig. 39.4. The Cretaceous sediments are everywhere very thick and are almost entirely clastic. They probably have been studied most under the Arctic Coastal Plain and in outcrop in the Foothills Belt under the auspices of the U. S. Navy Department in Naval Petroleum Reserve No. 4. Cross sections A–A′ and B–B′ of Fig. 39.6 show the beds and structure there approximately as they are today and as restored to pre-Laramide time, respectively.

The discovery of the Barrow arch of Precambrian (?) rocks under the northern edge of the Coastal Plain was unexpected, but it points to a positive region there, and to a northern source of sediments in mid- and late Paleozoic times (Dutro, 1960).

In the Kuskokwim distict Cady *et al.* (1955) describe Upper Creta-

ceous strata in the immense thickness of 40,000 to 65,000 feet. The rocks are dark, interbedded shale and fine-grained graywacke. Breccia and conglomerate facies are present in a few localities. The record of crustal deformation and sedimentation is described as follows:

> . . . Sediments were eroded from emerged areas of the geanticlines and were carried by streams to the trough of the intervening Kuskokwim geosyncline, where scores of thousands of feet of sediments were deposited while subsidence continued, during latest Early Cretaceous and early and possibly middle Late Cretaceous time. The sediments were drawn from older rocks exposed in the geanticlines—phyllite, slate, quartzite, limestone, siltstone, chert, basalt, and andesite.
>
> The geanticlines, particularly the Aniak-Ruby geanticline continued to be uplifted rapidly during at least the early part of the Late Cretaceous time, and areas of sharp relief evidently appeared from which the older rocks were violently eroded and subjected to disintegration almost entirely mechanical. The disintegration products, chiefly angular silt and sand-size fragments, were transported fairly short distances to the Kuskokwim geosyncline. The submarine relief of the belt of the Kuskokwim geosyncline, like the subaerial relief of the geanticlines, was continually steepened in the early Late Cretaceous epoch, particularly along the borders of the trough. Sediments left by the streams in this marginal area formed loose, unconsolidated deposits that were continually and repeatedly upset by the steepening of the trough borders, and slid down the submarine slopes of the trough. Part of the silt and sand involved in the slides became incorporated in turbidity currents of high density and were distributed in the otherwise unagitated water below wavebase. The sediments of the slides and of the turbidity currents came to rest to form the interbedded graywacke and shale of the Kuskokwim group. The graywacke beds formed at the time of sliding, and are possibly related to turbidity currents capable of transporting the sand-size particles. The latter settled at depths at which the currents were checked by seawater of equal density. Shale beds were laid down in more quiet intervals of settling. Beds of graywacke, many of which are as much as two feet thick, were probably formed in a very short time by this process, an instant of time in the geologic sense (Cady *et al.* 1955).

In the Alaskan peninsula 800 feet of Cretaceous arenaceous limestone occur in the Herenden Bay area, and 1175 feet of marine limestone, sandstone, shale, and conglomerate are noted on the southeast flank of the Talkeetna Mountains (Miller, 1959). These are part of the Matanuska basin. See Fig. 39.2.

Earliest Tertiary deposits occur in the Matanuska Valley and consist of shale, sandstone, conglomerate, and coal. They have a maximum thickness of 7000 feet and are tentatively assigned to the Paleocene. They are considered by Payne (1955) to represent the closing phase of sedimentation in the Matanuska basin.

MESOZOIC AND CENOZOIC OROGENIES

Belts of Orogeny

The eugeosyncline of southern Alaska during the Paleozoic and Mesozoic eras bespeaks almost constant orogeny. The platform region to the north was involved principally in Mid-Cretaceous and Early Tertiary orogeny. The southern margin of the eugeosyncline was involved in deformation during the Cenozoic, and thus a parallel with the western United States is at hand, for we have to deal with the Laramide belt on the north, the Nevadan belt in the southern half, and the Coast Range belt along the southern margin.

Nevadan Orogenic Belt

The Nevadan belt (Fig. 39.2) is characterized by Paleozoic and Mesozoic eugeosynclinal strata, by their intense deformation and low-grade metamorphism over large areas, by voluminous and numerous batholithic intrusions, and by the fact that the climatic orogenic events took place in the latter half of the Mesozoic. Three phases of Mid- and Late Jurassic orogeny are noted by unconformities in the Jurassic sequence in the Matanuska basin, and a fourth phase in the earliest Cretaceous (Miller, 1959). Intrusive activity began in Mid-Jurassic and continued through Late Jurassic in the Talkeetna geanticline. The major and intense deformation of the Alaska and Seymore basins occurred in late Early Cretaceous time (late Neocomian and Aptian). It was accompanied by the intrusion of the major batholiths there.

Jurassic orogeny is obscure in central Alaska but strata of pre-Albian age (late Early Cretaceous) are strongly deformed and considerably intruded. The batholiths may be Jurassic as well as Early Cretaceous.

The Kuskokwim group of Late Cretaceous age in the Kushokwim basin and similar strata in the Yukon basin (Fig. 39.2) were strongly folded,

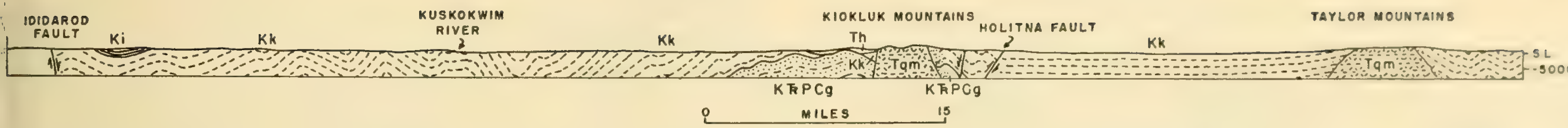


Fig. 39.7. Cross section of central Kuskokwim region, Alaska. After Cady *et al.*, 1955. Th, Holkuk basalt; Ki, Ididarod basalt; Kk, Kuskokwim group; KŦPC$_g$, Gemuk group.

probably in late Paleocene time. See structure of Fig. 39.7. According to Cady *et al.* (1955);

> . . . Late in Late Cretaceous time the deposits in the geosyncline were uplifted slightly above sea level, and the lava flows of the Iditarod basalt spread out over the uppermost strata of the Kuskokwim group.
>
> The geanticlinal tracts moved closer together in earliest Tertiary time, probably because the more rigid continental platform and Pacific Ocean floor approached one another and decreased the width of the mobile belt. The geosynclinal accumulations of the Kuskokwim group, which were structurally less competent than the geanticlines, were as a result thrown into folds that were draped around the margins of the geanticlines, and were also grouped into rather extensive anticlinorial uplifts, such as the Gemuk anticlinorium, which includes an upbuckled portion of the floor of the Kuskokwim geosyncline. Biotite basalt sills and dikes and albite rhyolite sheets, sills, and dikes, partly concordant with the enclosing formations, were intruded in the geosynclinal rocks and underlying strata near the close of folding.

Nonmarine Early Tertiary rocks, presumably of Eocene age, occur as erosional remnants on the Nevadan complex in several areas of central Alaska, particularly in the Healy basin (Fig. 39.2), north of the Alaska Range. They consist of claystone, sandstone, conglomerate, and lignite up to 5000 feet thick. Sedimentary rocks of this age are believed to have been deposited extensively in what are now the alluvium-floored lowland basins. These Eocene sediments were gently, and locally strongly, deformed in Oligocene or early Miocene time.

Strong uplift occurred lastly at the close of the Tertiary and during the Quaternary to produce the high mountain ranges and upland areas of central Alaska.

Coast Range Orogenic Belt

The Coastal Range orogenic belt as here defined is much like the Central and Northern Coast Ranges of California, inasmuch as the bedrock geology is the Nevadan complex with deformed Tertiary beds superposed.

The belt shown on Fig. 39.2 is widest in the Cook Inlet and Prince William Sound region where it includes the Chugach and Kenai Mountains. Three areas of Tertiary rocks are recognized, the Gulf of Alaska Tertiary province, the Cook Inlet Tertiary province, and the Aniakchak Tertiary province.

Gulf of Alaska Tertiary Province

Stratigraphy. The Yakataga basin of Tertiary deposits is an arcuate lowland and foothills belt. The province borders the Gulf of Alaska from the Copper River delta 300 miles southeastward to Icy Point, and extends inland up to 40 miles to include the southern front of the Chugach and St. Elias ranges. Although generally lowlands, the Gulf of Alaska Tertiary province includes groups of hills and unnamed mountains in the Katalla district up to 5000 feet above sea level, the Robinson Mountains in the Yakataga district rising to 9000 feet, the Chaix and Samovar Hills along the north margin of the Malaspina Glacier to 6000 feet, and a ridge in the Lituya district up to 3500 feet. Elevations above 1500 feet are covered by permanent snow fields and glaciers.

Typical sections of the Tertiary rocks are given in Fig. 39.8.

> Three major subdivisions of Tertiary rocks are recognized on the basis of gross lithologic characteristics and fossil evidence. These units are believed to correspond to major changes in the depositional environment of the Yakataga geosyncline.
>
> The oldest unit, of Eocene and possibly early Oligocene age, consists predominantly of interbedded or intertonguing nonmarine coal-bearing strata and shallow marine or brackish water strata. Fossil plants and marine invertebrates in this unit are regarded as indicating subtropical to temperate climate on land and tropical to warm-temperate marine environment. This

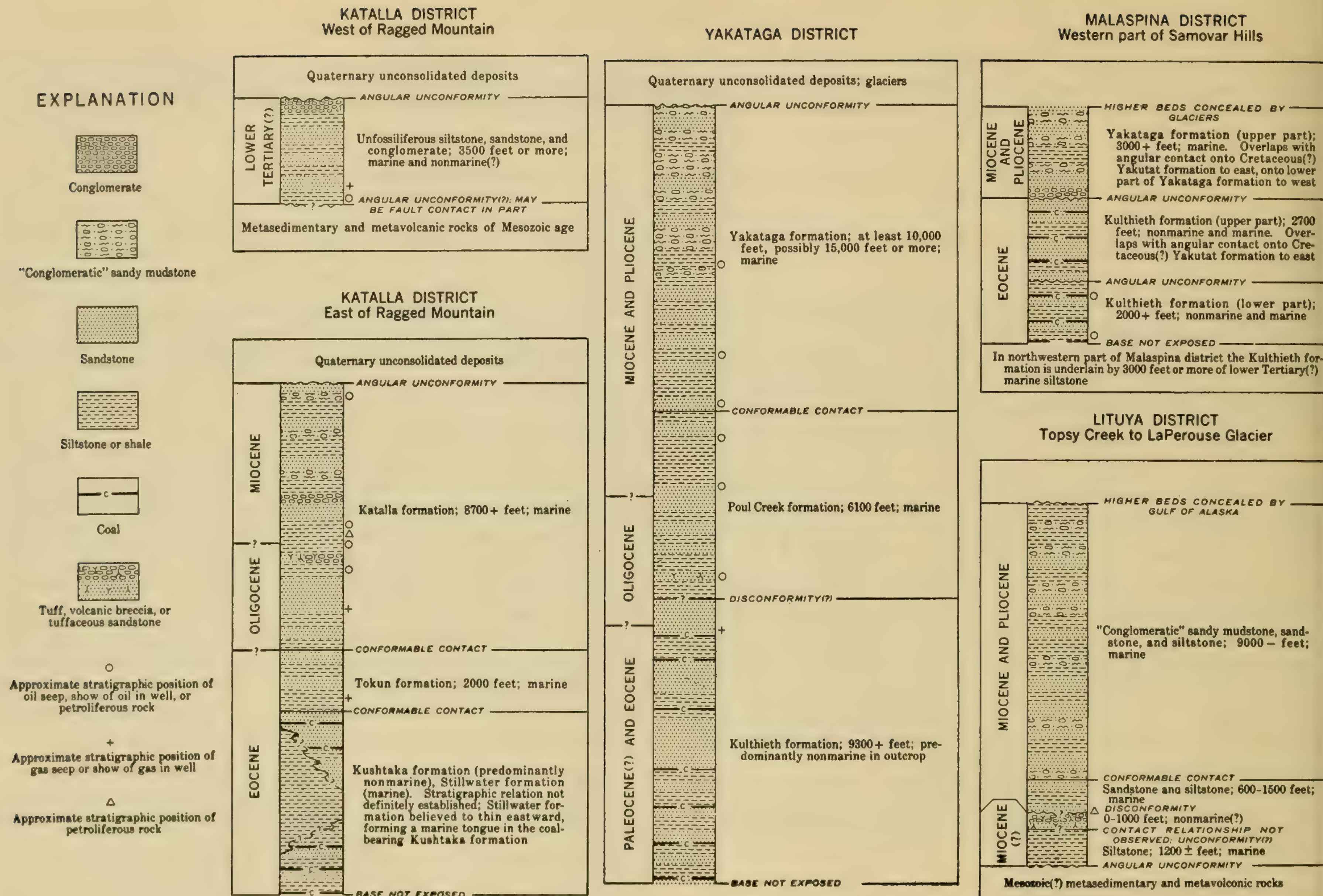


Fig. 39.8. Representative stratigraphic sections of the Tertiary sequences exposed in the Gulf of Alaska. Tertiary province. Reproduced from Miller *et al.*, 1959.

unit includes the Kushtaka, Stillwater, and Tokun formations in the Katalla district, and the Kultheith formation in the Yakataga and Malaspina districts. It is not represented in the exposed Tertiary sequence of the Lituya district.

The middle unit, formed in middle Oligocene to approximately middle Miocene time, is characterized by massive concretionary mudstone and siltstone, believed to have been deposited in moderately deep water, in part in a reducing environment. Local volcanic activity is indicated by interbedded marine tuff and agglomerate. This unit is highly organic at some places, and many of the known indications of petroleum in the Katalla and Yakataga districts are associated with it. The unit includes the lower and middle parts of the Katalla formation in the Katalla district, the Poul Creek formation in the Yakataga district, and the basal part of the exposed Tertiary sequence in the Lituya district. It is absent in the exposed Tertiary sequence in the Malaspina district, where the early and late Tertiary units are in unconformable contact.

The youngest unit, deposited during the time interval from middle or late Miocene to late Pliocene or possibly earliest Pleistocene, consists of shallow marine sandstone and siltstone interbedded with marine tillite ("conglomeratic" sandy mudstone). The marine invertebrate fauna, on the whole, indicates considerably colder water than in earlier Tertiary time, and the marine glacial deposits indicate rigorous glaciation of adjacent land areas. This unit is represented by the upper part of the Katalla formation in the Katalla district, by the Yakataga formation in the Yakataga and Malaspina districts, by the upper part of the unnamed sequence in the Lituya district, and by strata exposed on Middleton Island, a small island in the Gulf of Alaska 80 miles southwest of Cordova (D. J. Miller, 1959).

Structure. The structure and orogenic history of the Gulf of Alaska Tertiary Province is described by D. J. Miller (1959) as follows (Fig. 39.9):

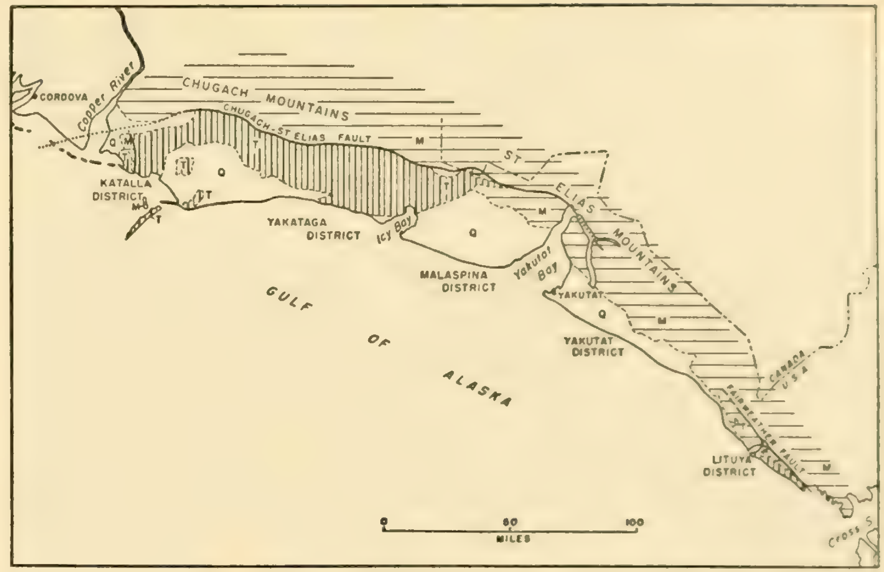


Fig. 39.9. Generalized geologic map of Gulf of Alaska Tertiary Province, after Miller *et al.*, 1959. Q, lowland area covered by ice or unconsolidated deposits of Quaternary age; possible underlain by sedimentary rocks of Tertiary age. T, sedimentary rocks of Tertiary age. M, metamorphosed sedimentary rocks and volcanic rocks of Mesozoic and older(?) age.

In late Tertiary or early Pleistocene time the Chugach–St. Elias Mountain chain was uplifted along an arcuate northward-dipping fault system, and the bordering belt of Tertiary sedimentary rocks was folded and displaced along many high-angle thrust faults. The largest of these faults, the Chugach–St. Elias fault, has been traced along the southern front of the Chugach and St. Elias mountains from the delta of the Copper River to Yakutat Bay, a distance of 180 miles. This fault, which dips 30°–60°N., is estimated to have a stratigraphic throw of not less than 10,000 feet. In the Lituya district the Fairweather fault, lying in a great trench at the base of the Fairweather Range, bounds the Tertiary province.

The major thrust faults and grain of folding in the Tertiary rocks in general parallel the trend of the bordering fault system along the Chugach–St. Elias front; the intensity of folding and magnitude of displacement along faults increases toward the mountain front. Transverse trends in the western part of the Katalla district apparently are related to the northward-trending Ragged Mountain fault that exposes the pre-Tertiary basement rocks. In the Katalla district the folds are typically of small amplitude, tightly compressed, and asymmetric, the axial planes being inclined to the west or north.

In the Yakataga district three belts of differing structural pattern are recognized: In the belt nearest the Chugach–St. Elias fault the Tertiary rocks show intense minor folding with much overturning, and are displaced along many northward-dipping high-angle thrust faults, which in general are subparallel to the axial planes of the folds. In the intermediate belt the folds are of small amplitude but relatively long, and are less tightly compressed and more widely spaced. The belt nearest the coast is characterized by broad synclines and narrow, tightly pinched, asymmetric, longitudinally faulted anticlines.

In the Malaspina district, faulting and uplift predominated over folding during the late Cenozoic orogeny, for the youngest Tertiary strata are only broadly folded or gently tilted. At least two earlier stages of deformation and uplift within the Tertiary period are recorded by angular unconformities within the Kulthieth and Yakataga formations, and by overlap of the upper

part of the Yakataga formation on early Tertiary and pre-Tertiary rocks.

Near Lituya Bay in the Lituya district the narrow belt of Tertiary rocks is folded into a shallow syncline and a strongly asymmetric anticline. These folds pass to the southeast into a seaward-facing homocline which, at Icy Point, is overturned. Upper Tertiary rocks in the outlier in the northern part of the Lituya district form a broad syncline trending northwest.

Cook Inlet Tertiary Province

The Cook Inlet Tertiary province includes the Cook Inlet lowland and the lower part of the Susitna River valley. About 75 miles to the east of the lower Susitna River basin and separated from it by the Talkeetna Mountains, is the Copper River Tertiary and Quaternary basin. The basins are floored extensively with Quaternary deposits, and these are believed to cover Tertiary beds which crop out mostly in marginal areas. See map, Fig. 39.1.

Stratigraphy. The chief display is a coal-bearing series of nonmarine clastics. In the Kenai lowland the strata have been named the Kenai formation. They consist of partly indurated sand, silt, clay with thin conglomerate lenses and many thin beds of sub-bituminous coal or lignite, and have a thickness of at least 4700 feet. The formation is presumed to be Eocene and to rest unconformably on the deformed Mesozoic rocks.

At a locality on the northwest margin of the Cook Inlet province, 900 feet of clay, sand, and gravel, presumed to be Eocene, rests with angular unconformity on highly deformed slate and graywacke of Mesozoic age. The unconsolidated beds are overlain, apparently conformably, by 500–1100 feet of coarse gravel, possibly of Oligocene or younger age.

Structure. The Tertiary beds of the Cook Inlet province are not as much deformed as those of the Gulf of Alaska province. For the most part they are nearly flat or only gently tilted or folded. In some marginal areas dips up to 60 degrees have been observed.

Aniakchak Tertiary Province

A Tertiary area, here called the Aniakchak Tertiary province, composes the southwestern half of the Alaska Peninsula. The Upper Jurassic and Cretaceous rocks are overlain with minor unconformity by Early Tertiary nonmarine, coal-bearing arkosic sandstones and shales and much fragmental volcanic material interbedded with flows. These rocks are presumed to underlie much of the Shelikof Strait depression. See Fig. 39.1.

Marine strata of Eocene, Miocene, and Pliocene (?) age are exposed in the Herendeen Bay area and Shumagin Islands.

The Early Tertiary strata of the Alaska Peninsula are in general gently tilted or folded. Several well-defined anticlines with flank dips of 5 to 45 degrees have been mapped. One is 30 miles long.

Laramide Orogenic Belt

The Laramide orogenic belt lies north of and adjacent to the Nevadan. It is made up of two parts, the Foothills or gently deformed belt, and the main or strongly deformed. In contrast to the Nevadan belt, the Laramide involves the Paleozoic platform-type sediments, as well as Mesozoic sediments mostly of miogeosynclinal nature. Also, intrusive masses are few and not so large as in the Nevadan. The line drawn on Fig. 39.2 separating the Nevadan from the Laramide was determined mostly from the distribution of late Mesozoic intrusions, viz., most of the intrusions lie south of the line. Included in the Laramide belt, accordingly, are the Yukon basin, Seward uplift, Hogatza arch, Kobuk basin, Brooks Range geanticline, and the Arctic Foothills belt.

Brooks Range Geanticline. The northern limb of the Brooks Range geanticline consists of slightly metamorphosed Devonian and Carboniferous rocks. Dark clastic rocks of the Sadlerochit formation (Permian and Early Triassic) generally overlie the lighter carbonate rocks of the Lisburne group (Mississippian) and form conspicuous hogbacks along the northern edge of the range. The structure of the northern half of the geanticline is one of folds and thrusts.

The southern limb consists of early Paleozoic metamorphic rocks and Silurian limestone. Tight folds and thrust faults toward the north repeat the formations in numerous subparallel belts (D. J. Miller, 1959).

At least 10,000 feet, and perhaps 15,000 feet, of Devonian and Carboniferous sedimentary rocks including much limestone were deposited in a Paleozoic basin in the area of the present Brooks Range. Most of the

clastic materials were probably derived from an uplifted shield north of the present land area, according to Miller. Permian rocks in the western Romanzof Mountains area become coarser toward the north.

The Brooks Range geanticline began to rise in Jurassic time (Miller, 1959). In one place mafic and ultramafic intrusions were emplaced in Late Jurassic time. The main phase of orogeny occurred in Aptian time (late Early Cretaceous) when the metamorphism of the rocks was accomplished under deep burial, and an east-west structural pattern took form. Uplift occurred throughout Late Cretaceous time and much debris was shed to the Colville basin. A late (?) Paleocene phase of deformation possibly resulted in the thrust faults, but these may have formed earlier, and the east-west structural grain was intensified. Peneplanation, and Quaternary uplift followed.

Romanzof Uplift. The Romanzof uplift appears as a northern bulge of the Brooks Range. Fold axes plunge westward in the Canning River area, and strata primarily of Carboniferous, Devonian, and possibly Precambrian ages are exposed. Mesozoic rocks are preserved in certain structural depressions. The general uplift started in mid-Cretaceous, or possibly earlier, and continued in uplift during the Tertiary.

Tigara Uplift. A small area of complexly folded and faulted rocks of Devonian, Carboniferous, and early Mesozoic age is exposed along the coast line between Cape Lisburne and Point Hope, north of the De Long Mountains. These older rocks rise from the Southern Foothills belt (Index map of Fig. 39.6) and are called the Tigara uplift. It must be a more extensive feature under the shallow water to the west.

Seward Uplift. The Seward peninsula is made up largely of deformed Paleozoic rocks with Cretaceous intrusions and three large areas of Tertiary volcanic rocks. The most extensive area of Ordovician rocks in Alaska is in the western part of Seward peninsula. The rock is dominantly limestone, and the beds have been cast into broad open folds and show little effects of dynamic metamorphism. Their exact thickness is not known but at least 5000 exist (Smith, 1939).

There are also large thicknesses of Silurian, Devonian, and Carboniferous limestones on Seward peninsula, but identities, correlations, and thicknesses are not yet well known. Although the Ordovician strata of the western part of the peninsula are only gently folded, the strata of other areas are intensely deformed.

According to Payne (1955 and 1959) the dominant structural grain is east-west and represents Early Cretaceous and possibly Late (post-Portlandian) Jurassic phases of orogeny. Basic intrusions came in first and then a number of large stocks or small batholiths of more acidic rocks. The granitic intrusions with accompanying local metamorphism and mineralization occurred probably in Aptian time. The peninsula thereafter remained mostly emergent and furnished sediments to adjacent basins, particularly the Yukon. In early Tertiary time a second episode of deformation produced a north to northeast grain superimposed on the older east-west grain. Faulting was prominent.

During the Tertiary, erosion was extensive but the peninsula remained broadly above sea level. Considerable volcanism occurred in late Cenozoic time and resulted in blankets of extrusive rocks over the deformed Paleozoic complex.

As portrayed on the map of Fig. 39.2, the Seward uplift included not only the Seward peninsula but an approximately circular region under the shallow water of Norton Sound and the Bering Sea. Although a positive area in Mesozoic and Cenozoic times it is considered part of the general Laramide belt. The phase of major deformation and intrusions, here as in the Brooks Range, appears to have been late Early Cretaceous, and orogeny of this age is generally considered to be pre-Laramide in the Rocky Mountains of the western United States. However, as previously explained, the Laramide belt is defined by physical characteristics as well as time of orogeny and a phase of deformation earlier than Late Cretaceous is a normal attribute of the Laramide belt.

Arctic Foothills Belt

A belt of "plateaus standing at different elevations" (Mertie, 1930) lies north of the Brooks Range, and much work incident to the exploration of Naval Petroleum Reserve No. 4 has established clearly that this is a foothills belt, both topographically and structurally. It is subdivided into two sections, the southern foothills and the northern foothills. See Fig. 39.6.

Southern Foothills. The southern foothills are characterized by isolated, irregular hills and ridges of sandstone, limestone, and chert which rise above low shale areas of little relief. This section has the structural complexity of the Brooks Range but differs in being composed of less resistant rocks, including a great thickness of shale. Ridges and hilltops are at altitudes of 2500 to 3500 feet and rise 1000 to 2000 feet above the surrounding plains. The southern foothills are readily traversable by such vehicles as the weasel but not so easily by boat, plane, or foot. Lakes suitable for landings by small float planes (1 to 2 passengers) are not abundant, and only a few lakes such as Noluk and Liberator are suitable for larger float planes (3 to 6 passengers). The flat areas between the hills or along ridgetops are ideally suited to the use of tracked vehicles.

Northern Foothills Section. The northern foothills section differs from the southern section in having more regular topography, including persistent ridges and elongate mesas that reflect a simpler structure of Appalachian-type folds, with minor cross faults, and a few major overthrusts. Anticlines are commonly asymmetric with steeper limbs on the north.

TERTIARY VOLCANIC ROCKS

Volcanics of the Coast Ranges

The Wrangell Mountains (Figs. 37.1 and 39.1) consist of a major Quaternary stratovolcanic accumulation. At least four major centers of eruption form a cluster of majestic peaks, namely, Mt. Wrangell (14,000 feet), Mt. Drum (12,000 feet), Mt. Sanford (16,210 feet), and Mt. Blackburn (16,140 feet).

Of these only the first-named has been seen "smoking." Apparently volcanism in this region did not begin until some time after an early Tertiary plain of erosion had been formed, uplifted, and somewhat dissected. Since that time there has been almost unceasing volcanic activity in different parts of the area, during which the present huge agglomeration of flows, breccias, and tuffs has accumulated. Most of these rocks are porphyries of medium coarseness and light or dark-gray color. In composition the usual type is a hypersthene or hornblende andesite, but more basic or more acidic phases range from basalt to dacite. The color of these rocks also shows a considerable variation from the type, as brick-red, pink, lavender, brown, and greenish tones are by no means rare. The eastern limit of the lavas in the Copper River region that may be correlated with the Wrangell lava is in the mountains adjacent to Skolai Pass, where they cap many of the highland areas and unconformably overlie Paleozoic and younger sedimentary rocks. That the lavas in this area are correlative with the older members of this volcanic series seems clearly indicated by the extensive dissection they have undergone, whereby the deep valleys of Skolai Creek and the Nizina River and Nizina Glacier have been deeply trenched through them. None of these Tertiary-Recent lavas shows evidence of marked deformation after they were poured out. The thickness of the lava series differs considerably in different places, and no measurements are available that disclose the total thickness of these beds in the heart of the range. Partial sections have shown more than 4000 feet of these volcanic rocks near Regal Glacier, in the Nizina Valley (Smith, 1939).

Cook Inlet–Susitna Field

Overlying the sedimentary coal-bearing and associated rocks in the Matanuska area and extending both eastward into the Nelchina area and northward into the Talkeetna Mountains is a series of andesitic basalt flows with intercalated tuffs. They are nearly horizontal and at least 1000 feet thick. They are deeply dissected and form cappings of the highlands. In the Nelchina area certain rhyolites appear. The series is thought to be late Eocene to Miocene in age (Smith, 1939).

Tertiary volcanic rocks are widespread in the Nevadan belt and only a few examples will be mentioned.

Volcanics of the Nevadan Orogenic Belt

In the Kuskokwim region the Lower (?) and Upper Cretaceous Kuskokwim group, is overlain disconformably by the Ididarod basalt, also of Late Cretaceous age. It is regarded as the first of a succession of volcanic rocks deposited in a continental environment. The Getmuna rhyolite group and the Holokuk basalt are early to mid-Cenozoic in age, and are separated from the older rocks by an angular unconformity. In late Cenozoic time the Waterboot basalt was erupted.

Intruded into the Kuskokwim sediments are a number of stocks of quartz monzonite, believed to be post-Holokuk basalt.

In the Yukon–Tanana region an older unit consists of rhyolite, dacite,

and andesite, with rhyolite the most abundant, and basalt practically absent. They are so widespread that they must have been erupted from several craters or from fissures. Mid-Tertiary intrusive granite rocks are probably later. In the younger group the acidic varieties occur, but basalt is common. Some of the more basic members have inclusions of ultrabasic composition. All the younger units are post-Miocene and in part Quaternary.

Farther down the Yukon in the Chandaler Valley and in the Koyukuk Valley volcanic rocks believed to correlate with the older unit of the Tanana region occur. These rocks are partly in the Laramide belt.

In the lower Yukon Valley volcanic rocks undoubtedly of several ages occur with the older presumably more acidic than the younger.

Volcanics of the Laramide Orogenic Belt

Other than the Tertiary volcanic rock occurrences in the Yukon Valley which are partly in the Laramide belt, the main eruptions have been in the Seward peninsula. Three large fields are shown on the tectonic map of D. J. Miller (1959). One has an area considerably more than 1000 square miles. Old flows occur but the bulk of the volcanic rock is typically Recent. The ropy surface is preserved, Quaternary gravels are covered, and stream drainages blocked. The sources are not evident, and perhaps the flows issued from fissures. The composition is basaltic (Smith, 1939).

ALEUTIAN VOLCANIC BELT

Kinds of Volcanoes

A great arc of volcanoes extends from Mount Spurr on Cook Inlet along the whole Alaska peninsula and the Aleutian Archipelago. See map, Fig. 39.10. This arc is 1500 miles long. Unfortunately, most of the volcanoes are situated in regions of sparse population little visited by outsiders, and therefore their grandeur is seldom seen. The highest stand 8000 to 11,000 feet above the sea and excel in beauty many of the venerated volcanoes of better-known lands. The Wrangell volcanic field and Mt. Edgecumbe extend the belt of active or recently active volcanoes another 1000 miles to the east and southeast.

Southward from the Mount Spurr group at the extreme northeastern limit of southwestern Alaska, the sites of Tertiary to Recent volcanism become increasingly evident until, at Mt. Veniaminof they include practically all the features of the bedrock. The lofty modern volcanoes that overshadow all the other topographic features are dominant in almost every landscape.

According to Coats (1950) there are at least 76 major volcanoes, active and extinct in the arc from Mt. Spurr to Buldir Island. Of these, 36 have been active since 1760. Seventeen calderas have been recognized. These are volcanic depressions, more or less circular, and over 1 mile in diameter. Of the 17 calderas the three largest are Fisher on Unimak Island which measures 10×11 miles, Aniakchak, 9.7×8.4 miles, and Veniaminof, 8.4 miles.

A number of volcanic domes have also been recognized. As defined, these are steep sided, viscous protrusions of lava forming a more or less dome-shaped mass around the vents.

> The older volcanoes of the arc seem to include both shield volcanoes, characterized by many relatively thin flows, with a small proportion of fragmental material, accumulated on slopes of low declivity, and stratovolcanoes or composite cones, made up both of flows and fragmental material, the slopes of which approach the angle of repose of the fragmental material. The major active volcanoes of the arc are without exception composite cones (Coates, 1950).

Petrographic Character

Smith (1939) summarizes the general petrography as follows:

> The composition of the lavas has in the main been fairly comparable with that of normal andesites, but more basic phases analogous to basalt and more acidic phases approaching rhyolite are by no means unknown.

Coates (1950) depicts them as follows:

> The volcanic rocks of the Aleutian arc range from olivine basalt to rhyolite. They include basalts characterized by olivine and andesites without olivine, in both of which hornblende and hypersthene occur separately or together. Relatively high percentages of conspicuous calcic plagioclase crystals and usually less conspicuous green augite characterize most of the rocks. Those that are comparatively rich in silica, such as dacites and rhyolites, are much

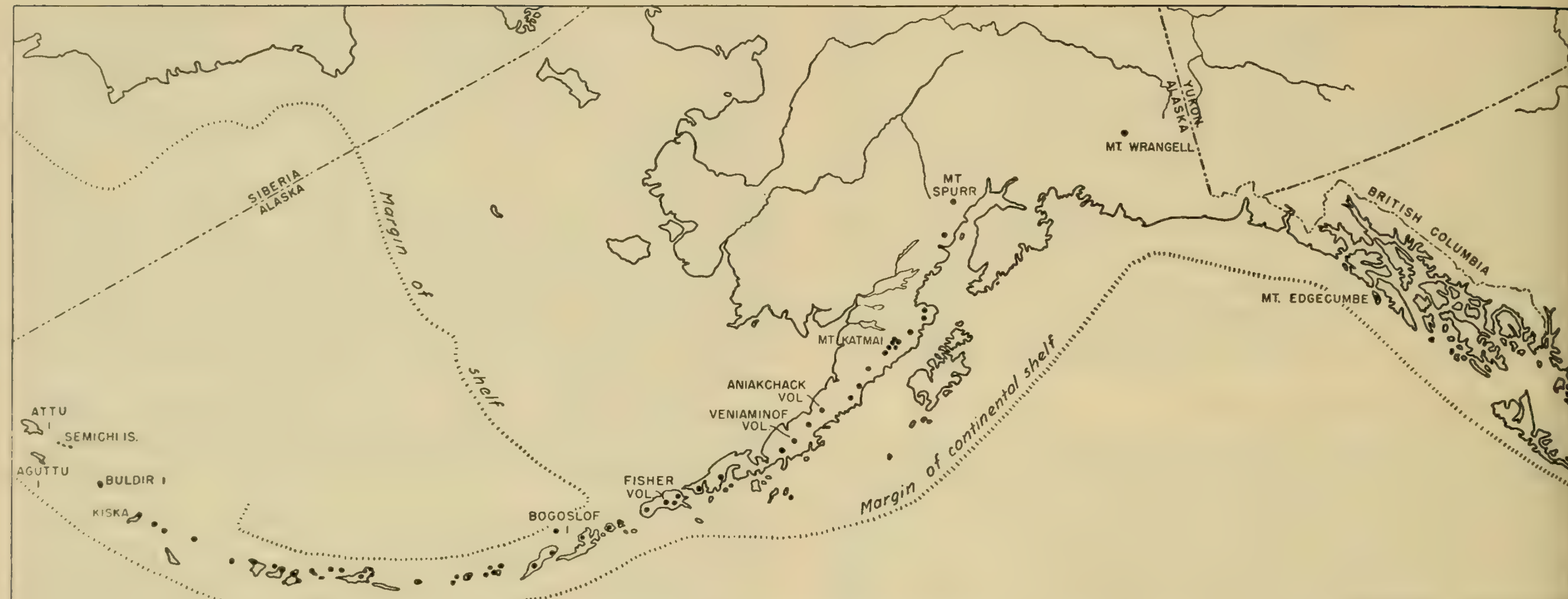


Fig. 39.10. Distribution of volcanoes in Alaska. Taken from Smith, 1939. Dots are active or recently active volcanoes.

less plentiful; most of them are present either as small bodies of highly glassy lava or as blankets of light-colored pumice.

Relation of Volcanism to Structure

In general, the volcanoes are superficial structures, built upon a basement of Tertiary and older rocks that is exposed at intervals throughout the length of the arc. The nature of the structures that have determined the position of the eruptive centers can be determined in few places. Some writers have thought that the line of volcanoes, because of its narrowness, represents the trace of a great thrust plane or fault, movement on which is thought to be responsible for the frequent earthquakes. In detail, the volcanic line does not form a perfectly simple arc, but consists of segments of different lengths; the included angles between adjacent segments may be as little as 140°. Certain volcanoes, like Bogoslof and Amak, lie some distance away from the main line, on the concave side of the arc. In the Aniakchak region, Knappen mapped a tension fault with an east-west trend, along or close to which several volcanic structures are alined; he considered that the site of the eruptive center was determined by the existence of the fault. It is probable that similar relationships exist elsewhere in the arc and that most of the volcanoes have had their sites determined by minor tensional fractures striking at an angle to the major overthrust zones. The distance of a volcano from the major active zone of movement is probably dependent upon the depth at which such a tensional fracture, originating in and limited to an overthrust block, taps eruptible magma (Coates, 1950).

Seismicity

The distribution of earthquake foci (Gutenberg and Richter, 1941, 1945) is such that the epicenters of shallow earthquakes tend to be south of the chain; those of intermediate-depth earthquakes (deeper than 60 kilometers) are in the islands of north of the chain. It seems probable that the general structural picture

of the Aleution Islands, when more information is available, will resemble that presented by Gutenberg and Richter for the structurally similar Japanese arc (1941). The distribution of deep and intermediate earthquake foci will probably fall along an active zone or surface, which will be shown to reach the surface of the crust along the northern slope of the Aleutian Trench and to dip northward at a moderate angle (Coates, 1950).

Age of Aleutian Arc

The southern part of Kiska and the nearby islands of Attu, Agattu, and the Semichis at the west end of the Aleutian arc lack the young stratovolcanoes characteristic of the central and eastern islands. Instead, they are composed of pre-middle Tertiary rocks and subordinate amounts of Upper Tertiary coarse clastic sediments and subaerial lava flows. On Tanaga and Oglinga islands of the west-central Aleutians smoothly rounded boulders in gravel beds on a wave cut platform appear to represent the oldest rocks of the region. The rock types are hornfels, hornblende gneiss, slate, schist, granulite, granodiorite, biotite granite and hornblende granite. The bedrock from which the boulders were derived was not discovered. Judging from the lack of directional characters in the granites, they are presumed to be intrusive into the other metamorphic rocks.

A sequence of basalt flows, tuff-breccia, and agglomerate, intruded by large masses of gabbro and small masses of rhyolite, underlies most of the island of Adak and are known as the Finger Bay volcanics (Coates, 1956). These rocks have generally been greatly deformed and hydrothermally altered, although in no way metamorphosed like the metamorphic types in the boulders, which are therefore considered older.

A third sequence of basalt flows and tuffs, gray, hard argillite, and gray-green, coarse graywacke, seen on Attu and Shemya islands, has been intensely sheared and may be of intermediate age between the boulder rocks and the Finger Bay volcanics.

A plant fossil was found in the Finger Bay volcanics and identified as late Paleozoic in age (Coates, 1956). Therefore, the intermediate basalts and graywackes and the metamorphic and granitic rocks of the boulders are regarded as Paleozoic. Coates regards the gneiss, schist, granulite, granodiorite, and granite as continental types, and concludes, therefore, that a continental land area stood nearby from which the boulders were derived. This poses a difficult tectonic problem because the Aleutian Islands in this segment rise from a rather narrow welt which is flanked on each side, most probably, by oceanic crust. It seems possible to the writer that in the evolution of a great volcanic island arc from the oceanic crust that deep-seated metamorphism is possible, and that granitic type magmas can originate there by fractional crystallization. These acidic differentiates will not be large in volume such as those that arise in the master eugeosynclinal belts of the continental margin.

Although the evidence is preponderant that the Aleutian arc as we now see it is Cenozoic in age, we must recognize some much older aspects in its evolution. These are certainly not clear to us in their tectonic relations. As will be postulated under a later heading, the main tectonic elements of continental Alaska are believed to veer northwestward to the Anadyr Gulf and Chukotski peninsula of the Siberian mainland, holding within the confines of the Bering Sea shelf. See Fig. 39.2.

SIBERIAN TECTONIC CONNECTIONS

Aleutian Projection

Since the structures of the Alaska Range extend in a smooth curve into the Aleutian Range of the Alaska peninsula, and since the adjacent geanticlines and basins, including the Aleutian trench, project in the same direction, the natural inference has been that the Nevadan and Coast Range orogenic belts run out to sea and mostly die out abruptly or continue as a single geanticline concealed by Tertiary volcanics. This is the main assumption of Carey (1958) in the presentation of his theory of the Alaskan orocline.

Anadyr–Chukotski Projection

In 1955 Payne showed on a tectonic map of Alaska the Colville basin and Brooks Range geanticline to project northwestward under the shallow waters of the shelf off Siberia toward Wrangell Island, and this view is reiterated by D. J. Miller (1959), who conceived the Seward and Tigara

uplifts to be part of a much larger uplift embracing the eastern end of the Chukotski peninsula.

Now, if the *Geologic Map of the U.S.S.R.* (1955) is consulted, the Chukotski peninsula and adjacent areas to the west are found to be made up of three geologic provinces, namely, (1) a deformed and considerably intruded Cretaceous basin on the north; (2) a Tertiary Coast Range province on the south; and (3) an intermediate Tertiary volcanic belt in which it appears that the volcanics rest mostly on the Cretaceous complex. See map, Fig. 39.2. The Cretaceous basin with its abundant Cretaceous volcanics and many batholiths and stocks seems similar to the central geanticline and adjacent basins of southwestern Alaska, and if tectonic connections on this basis are attempted several lines of evidence support the postulate.

The Coast Range orogenic belt is adjacent on the south in Siberia as in Alaska. St. Lawrence Island with its major intrusions appears to be Nevadan and falls within the projected Nevadan belt. See Fig. 39.2. The shallow water shelf of the Bering Sea will contain both belts of orogeny, and the outer margin of the shelf lies in the line of projection. By this theory an erosion surface of the orogenic belts would have been buried by the deltaic deposits of the Yukon and Kuskokwim rivers. The last evidence suggestive of the northwest bend of the Nevadan and Coast Range belts is the bathymetry of the shelf off the southeast side of the Alaska peninsula. If the map of Fig. 39.11 is referred to, it will be seen that the shelf is broad off Kodiak Island and westward to Unimak Island (Fig. 39.1), and then narrows so that hardly any shelf exists along the volcanic islands of the archipelago. The narrowing shelf margin projects almost exactly to the Bering Sea shelf margin, as if this is a major tectonic line. It may thus be imagined that this line marks the swing of the Coast Range belt toward the northwest and Anadyr Bay.

The Seward uplift then becomes a coigne around which the Nevadan belt wraps rather sharply.

The Aleutian Archipelago is here considered a welt or geanticline that has developed with customary curvature, volcanism, and trench from ocean basin crust, whereas the Nevadan and Coast Range belts are marginal to continental crust. The archipelago and the Coast Range belt have evolved probably simultaneously, although the archipelago is now very active while the Coast Range belt under the Bering Sea is quiescent.

Bering Land Bridge

With Nevadan, Laramide, and Coast Range belts extending from Alaska to the Anadyr–Chukotski region of Siberia there can be little doubt that land was continuous from one continent to the other many times from the beginning of the Cretaceous to the present.

Hopkins (1959) reports that if sea level were lowered 120 feet, only a channel 20 miles wide would remain. If lowered 150 to 180 feet an intercontinental land connection would be established via St. Lawrence Island and the Diomede Islands. If lowered 300 feet, presumably to the level during the maximum glaciation of the Wisconsin, Alaska and Siberia would be joined by an almost featureless plain nearly 1000 miles wide from the shrunken Bering Sea to the shore of the Arctic Ocean.

YUKON TERRITORY AND THE DISTRICT OF MACKENZIE

Geography

The principal mountains and rivers of Yukon Territory and the adjacent district of Mackenzie are shown on Fig. 39.1. The Selwyn Mountains form the major drainage divide, with the several tributaries of the Yukon River flowing to the west, and tributaries of the Mackenzie flowing to the east and north. The long arcuate Mackenzie and Franklin Mountains stand off to the northeast of the main Cordillera, with the Mackenzie River flowing between the two ranges. Several plains and plateaus in addition to those shown are recognized by various writers (Bostock, 1948; Martin, 1959), but the geographic nomenclature is not completely standardized.

Stratigraphy

Strata of every Paleozoic and Mesozoic system are present in the region as well as rocks of Precambrian and Tertiary age. Dominant rock types are as follows: Precambrian and Lower Cambrian, clastics; Middle and Upper Cambrian, Ordovician, and Silurian, carbonates, black shales, and

Fig. 39.11. Aleutian trench and Bering Sea, showing relation of broad shelf off the Aleutian peninsula to Alaskan-Siberian shelf. Reproduced from Murray, 1945.

bedded cherts; Middle Devonian, carbonates and shales; Upper Devonian, clastics; Mississippian, carbonates; Upper Pennsylvanian and Lower Permian, clastics; Triassic, shales and limestones; Jurassic and Cretaceous, clastics; and Tertiary, clastics (Hume, 1954; Martin, 1959). A cross section restored to the time of pre-Laramide deformation from west of the Barn Mountains to the Mackenzie delta is shown in Fig. 39.12. Several unconformities attest several times of crustal unrest with the formation of various basins and uplifts.

The first conspicuous disturbance occurred in the British and Barn Mountains area, probably during late Middle Devonian or Late Devonian time. Upper Devonian sediments derived from the uplift form a depositional body much like the Catskill delta (Martin, 1959). The area of uplift was probably mountainous for a while. Judging from the Upper Devonian clastics in the Brooks Range and the Barron arch under the Coastal Plain the uplift extended westward through northern Alaska as shown on Fig. 39.13 (Dutro, 1960).

The second conspicuous uplift occurred in Pennsylvanian time in the Richardson Mountains area. It seems to have proceeded in two impulses, one before Late Pennsylvanian time and one during the Late Pennsylvanian. The uplift was flanked by a complementary basin on the northwest. A Pennsylvanian basin exists also under the Arctic Foothills and Coastal Plain, whereas the rest of Alaska was emergent at the time, so that a partial and approximate view of Pennsylvanian conditions is shown in Fig. 39.13.

An Upper Triassic and Jurassic basin subsided in a general north-south direction in the Richardson Mountains area. Cretaceous beds are absent in northwestern Yukon toward the Brooks Range, but reach considerable thickness in the Mackenzie Mountains near Norman Wells.

Laramide Orogeny

The present mountains, plateaus, and plains are the aftermath of Laramide deformation and some Cenozoic faulting, but the exact time of disturbance or the number of phases have not been well fixed. The Mackenzie and Franklin Mountains are foreland type, with gentle folds the dominant structure. High-angle faults are reported in places but no thrusts of typical Rocky Mountain fashion are known. The Franklin Mountains are reported as narrow, flat-topped anticlines, generally faulted on one side or both.

General structures of the Mackenzie Mountains and of other ranges in the region are shown in Fig. 39.14. As may be seen, folds dominate the structural types, but along the eastern side of the Richardson Mountains Jeletzky (1961) has mapped a fault pattern which he describes as follows (see Fig. 39.15):

> Major faults split the area into a number of irregularly shaped and structurally disconnected fault blocks, which differ strongly in the degree of structural complexity and age of their rocks.
>
> The structure of the area contrasts strongly with that of the central parts of Richardson Mountains, which is dominated by symmetrical, large, mostly

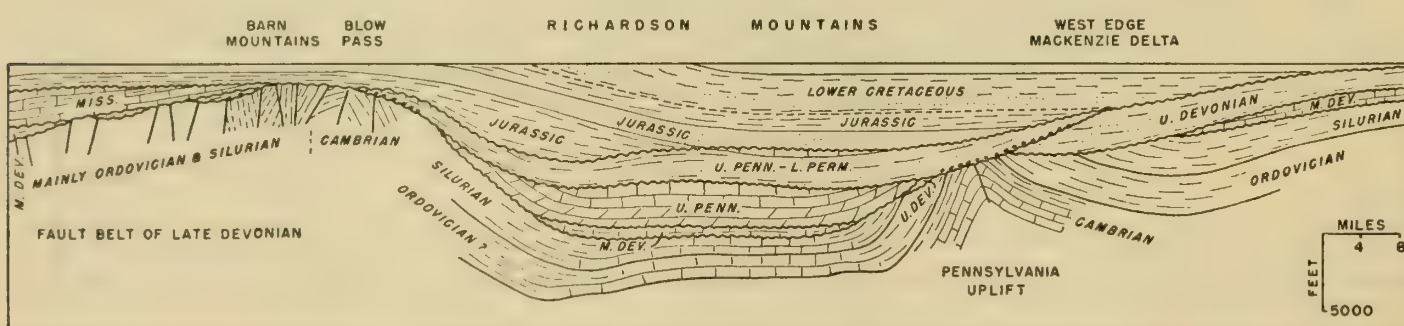


Fig. 39.12. Cross section from British Mountains to Mackenzie delta, restored to pre-Laramide time. After Martin, 1959.

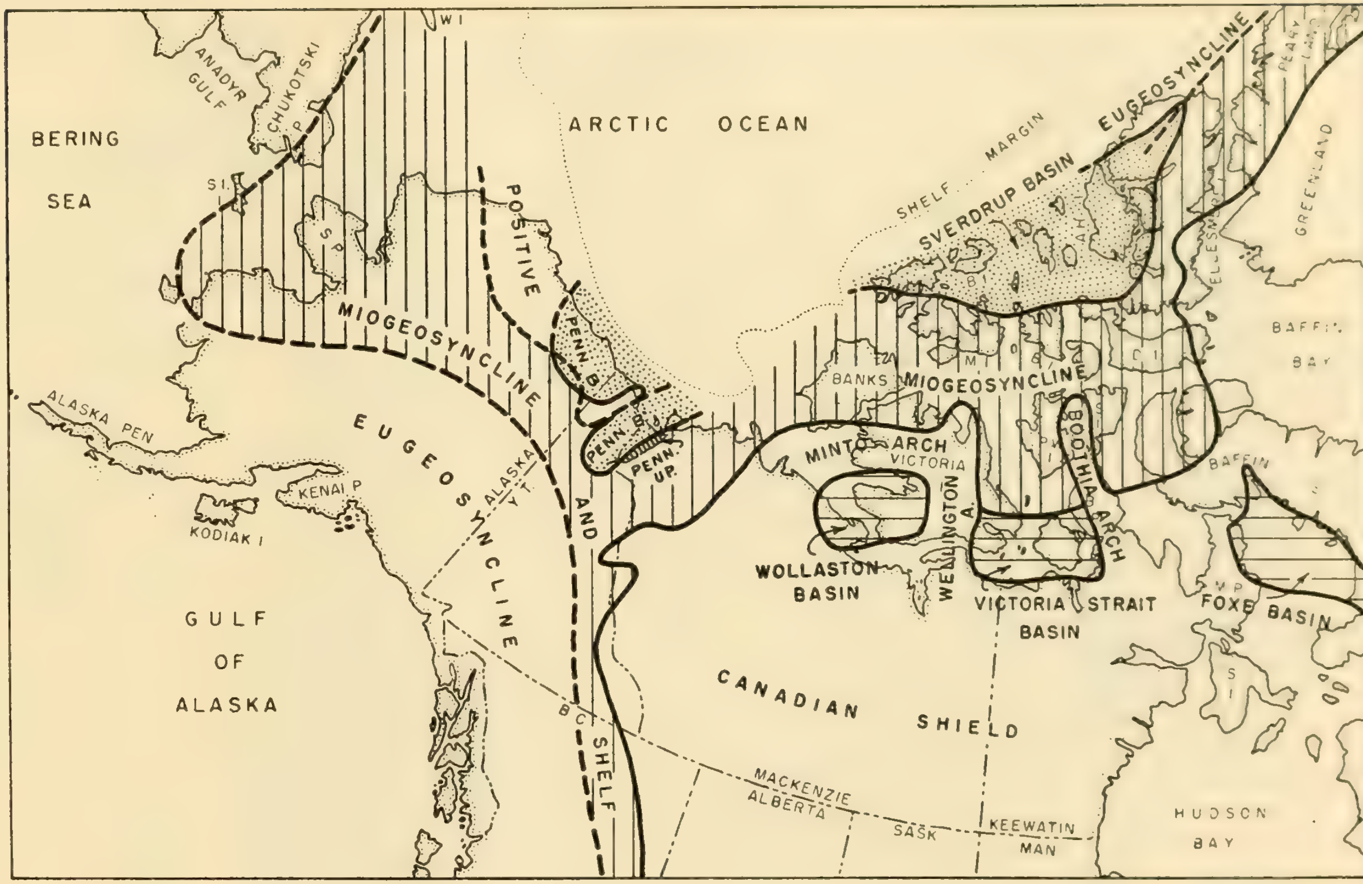


Fig. 39.13. Geosynclines, basins, and uplifts of Paleozoic age in Alaska and northern Canada. The fold belt of the Arctic Archipelago involves the miogeosyncline. The Wollaston, Victoria Straits, and Foxe basins are of the cratonic basin type. Most all Alaska appears to have been emergent in Pennsylvanian time. Ordovician and Silurian carbonate deposition was extensive from the Seward peninsula to the Yukon. P.P.I., Prince Patrick Island; B.I., Borden Island; A.H.I., Axel Heiberg Island; M.I., Melville Island; P.W.I., Prince of Wales Island; B.I., Bathurst Island; D.I., Devon Island; M.P., Melville peninsula; S.I., Southampton Island.

open folds, rather than by faults. The area appears, therefore, to form part of a separate structural zone, which separates the central parts of Richardson Mountains from the essentially stable belt situated further east.

Largest faults trend northerly and appear to be strike-slip faults. Folds are medium to small-sized, irregularly patterned, and commonly dome-like. Larger folds are strongly disrupted by faults and were apparently caused by an earlier orogenic phase. Smaller folds are subordinated to and were apparently caused by major faults. Both they and the major faults were, therefore, apparently caused by a later orogenic phase.

Hauterivian, late Aptian, early Albian, and late Albian or early Cenomanian (? at the Lower/Upper Cretaceous boundary) unconformities were observed in the area. Late Aptian unconformity is accompanied by a 5 to 10° angular discordance. Others are only recognizable because of smaller or larger transgressive overlaps.

The above unconformities were apparently caused largely by epeirogenic movements as no tectonic structures are known to be caused by them. The mid-Upper Cretaceous rocks of the area were, however, constantly involved in the major dislocations. The contemporary structures of the area were, therefore, caused largely or entirely by the post mid-Upper Cretaceous (? early Tertiary) orogenic movements.

A general description of the structures by Martin (1959) is informative.

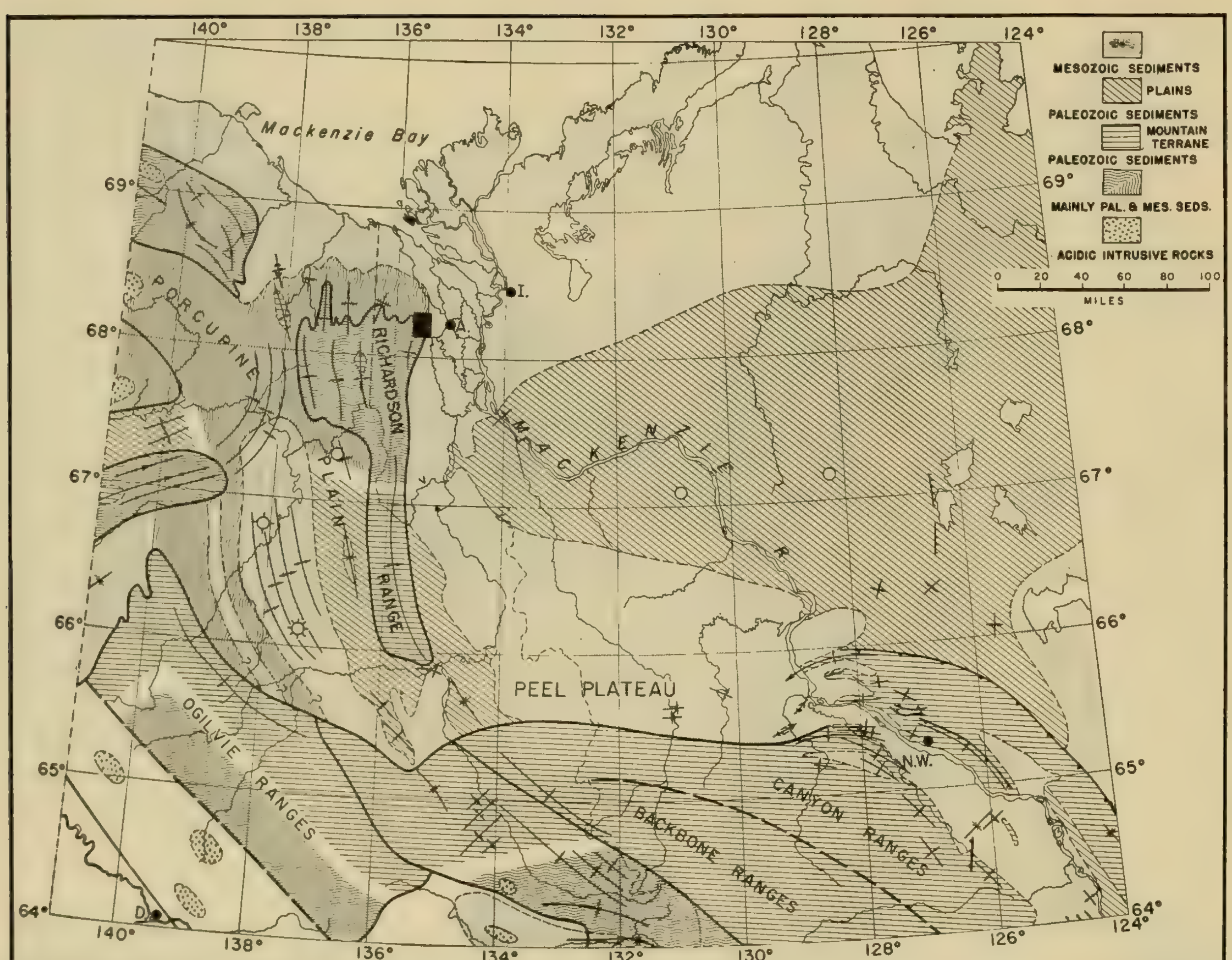


Fig. 39.14. Mountains and structural trends of the lower Mackenzie River region and northern Yukon. Map kindly supplied by P. E. Kent and W. A. C. Russell, British Petroleum, Ltd. Heavy lines surround mountainous areas. Dave Lord Ridge extends easterly from Alaskan border at 67 N. Lat. Canyon Ranges are northwest end of Mackenzie Range.

The fault structures (of the Ogilvie Mountains, Dave Lord Ridge, and northwestern Richardson Mountains) are of the same general type as those that form the Rocky Mountains of western Alberta, but a preliminary examination indicates that the stratigraphic displacement caused by individual faults is not as great as in the case of the Alberta Rockies. Faults in the Dave Lord Ridge area are irregular in trend and displacement, and do not result in the typical Canadian Rocky Mountain topography.

The anticline that forms the southern Richardson Mountains is of a type similar to the Wyoming Rockies uplifts, such as the Bighorn and Wind River mountains. The Franklin Mountains structures appear to be of the same type, but on a smaller scale.

The intrusion of the Old Crow Range batholith may have been in part responsible for Tertiary or late Mesozoic movements that took place in the British Mountains, and may have affected to some extent other structural patterns in the region.

Uplift of the coastal area following the retreat of Pleistocene glaciers is

demonstrated by the presence of raised beaches along the Arctic Coastal Plain and in the area west and southwest of the Mackenzie Delta.

Potassium-Argon Dates of Intrusives

The ages of several intrusives in the Yukon and the District of Mackenzie have been determined by Baagsgaard, Folinsbee, and Lipson (1961). The oldest date, 353 m.y., indicates an Acadian age. Two dates of 220 and 240 m.y. suggest late Paleozoic magmatic activity. Several dates ranging from 94 to 101 m.y. indicate intrusive activity in Mid-Cretaceous time or during the Nevadan orogeny. Figure 39.16 shows the position of the above intrusions and Fig. 39.17 shows the relation of Paleozoic orogenic belts and dated intrusions around the Arctic in Eurasia, Greenland, and northern Canada.

CENOZOIC TRENCHES AND FAULTS

Topographic Expression

The new Army Map Service Relief Quadrangles of Alaska, the Yukon, and northern British Columbia show strikingly five major linear topographic trends. Several smaller ones are also apparent. These linear features consist in part of trenches and in part of bold mountain escarpments, but the continuity of one with the other cannot be doubted. The Rocky Mountain, Tintina, and Shakwak trenches mentioned in Chapter 37 are especially clear on the maps. Some have been partially described in the literature and mentioned on previous pages of this chapter. The major alignments are so striking and the geomorphic provinces on either side in places so distinct that the writer is prone to consider them major, if not the most important, structural features of central and southern Alaska and adjacent regions. They are emphasized by bold lines on the map of Fig. 39.2.

Tintina–Rocky Mountain Trench Fault Zone

It is fairly evident that the Rocky Mountain trench projects to the Tintina trench, and thence to the south side of the Yukon Flats in

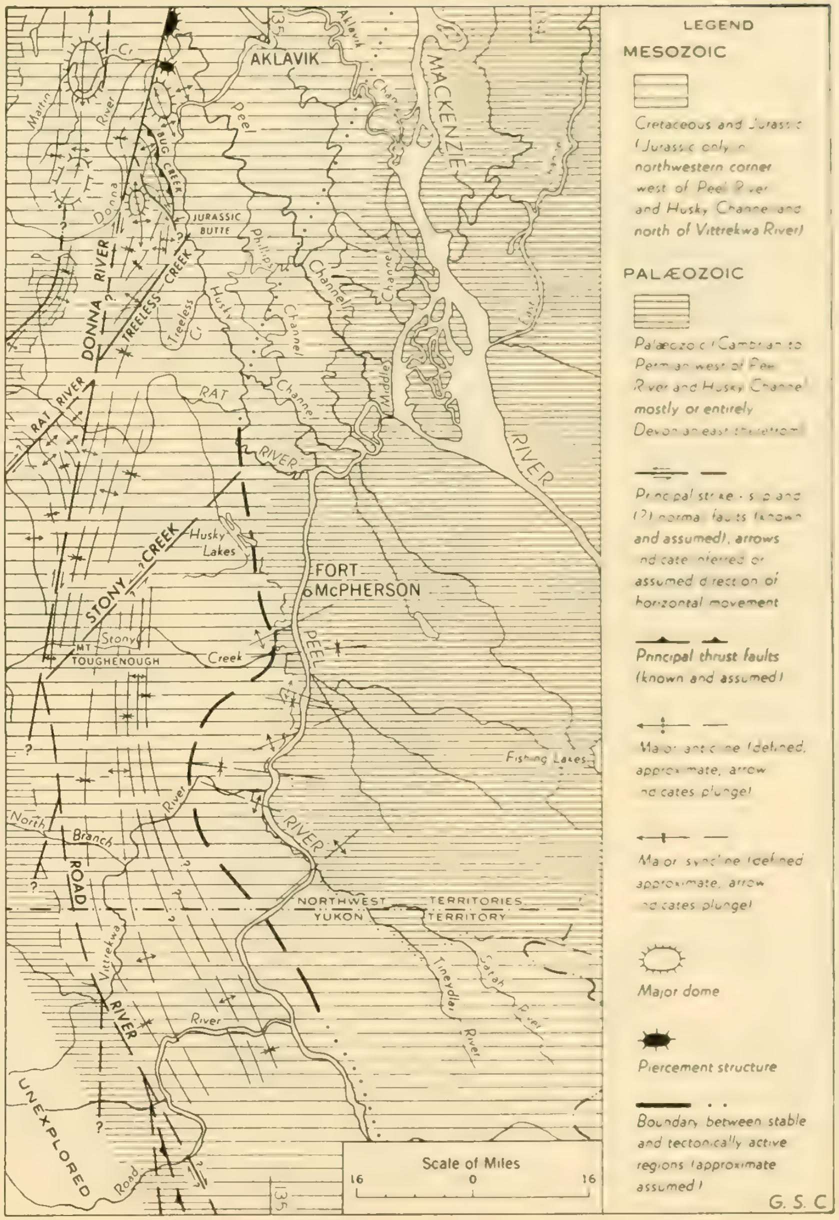


Fig. 39.15. Structures of the southern Richardson Mountains. Reproduced from Jeletzky, 1961.

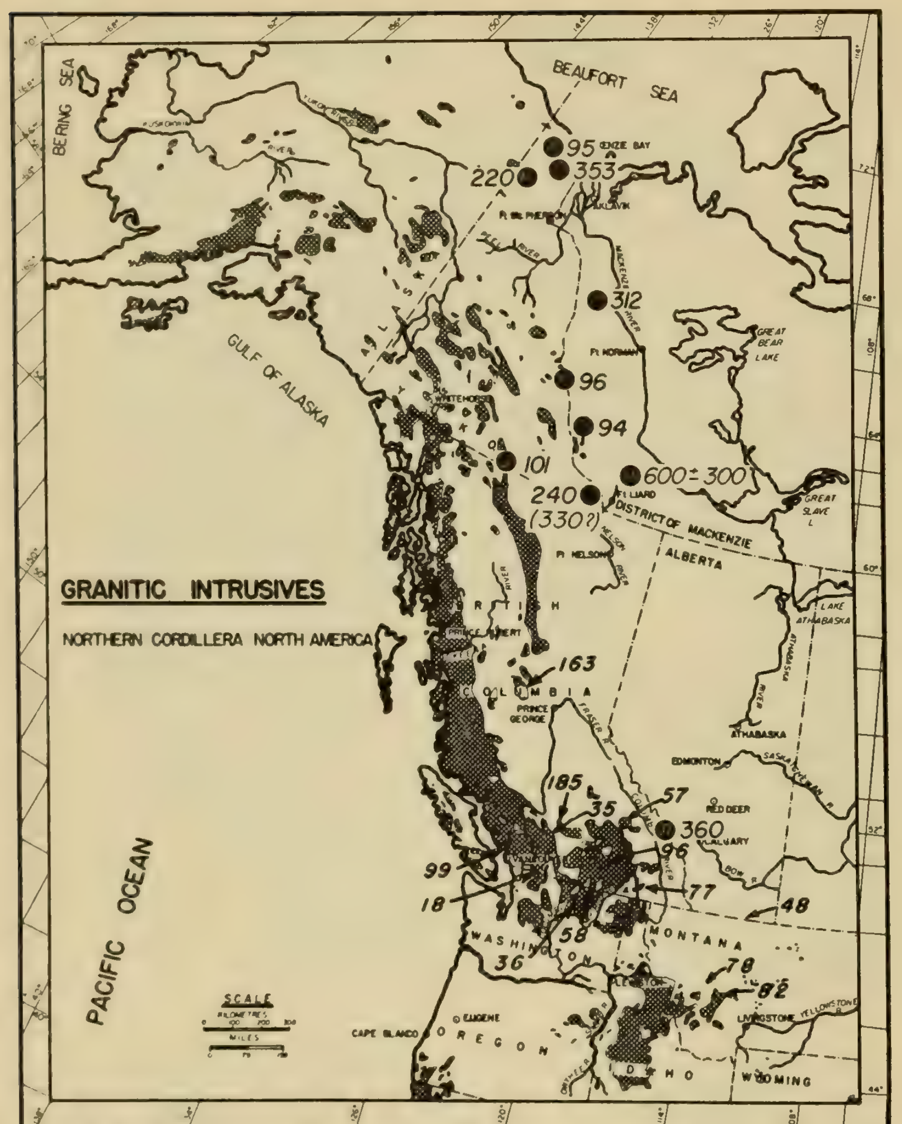


Fig. 39.16. Potassium-argon dates in British Columbia Mackenzie District and Yukon Territory. Map kindly supplied by R. E. Folinsbee. See Baadsgaard, Folinsbee and Lipson, 1961

Alaska, and down the Yukon River, possibly to the junction with the Tanana River. The bedrock is so much covered by alluvium from the Tanana down the Yukon Valley that the further course of the fault zone, if existent, is not evident.

Farewell–Shakwak Fault Zone

The most arresting alignment of valleys and mountain fronts starts on the northwest front of the Alaska Range (Mt. McKinley) and extends eastward as a trench through the southern part of the Alaska Range from Cantwell to Miller's Roadhouse, thence southeastward along the northeast front of the Mentasta and Nutzotin Mountains. It then crosses the border, follows along the Alaska Highway to Kluane Lake and to Dezadeash Lake where it jogs a bit to extend to the "Haines Cut-off" valley. It follows to the Lynn Canal. A branch may go out Chatham Straight, but the main fault appears to follow along Stephens Passage to Wrangell. The great fault zone has been named the Farewell in the Kuskokwim region and the Shakwak in the Yukon.

Mt. Logan Fault Zone

Not specifically pointed out in the literature as far as the writer knows is a major narrow topographic lineament just south of the Farewell–Shakwak zone. It is labeled "Mt. Logan Fault Zone" on the map of Fig. 39.2. Beginning on the southeast at Chatham Strait it proceeds as a trench along Icy Straight and Glacier Bay through the Mt. St. Elias Range to and along the Hubbard Glacier Valley. It then extends along the Logan Glacier Valley and the Chitina River. It thence passes a little south of the town of Chitina and down the Matanuska River Valley to Matanuska. It continues southeastward along the base of the mountains east of Anchorage to Kachemak Bay. This postulated fault zone is not as smoothly curved or linear as the others.

Chugach–St. Elias and Fairweather Faults

A great fault has been described in the Tertiary province of southern and southeastern Alaska and it is illustrated in Fig. 39.9. It is also vividly expressed on the relief maps.

Bruin Bay Fault

Along the northeast side of the Cook Inlet depression is the Bruin Bay Fault (D. J. Miller, 1959). Between it and the Mt. Logan fault is the Cook Inlet Tertiary province, described on a previous page.

Faults of the Kuskokwim Region

The Iditarod–Nixon fault, the Holitna fault, and the Farewell fault have been described on previous pages and are part of the great fault system.

Nature and Age of Faults

The faults of southwestern Alaska are undoubtedly high-angle faults, and Cady *et al.* apparently sees no evidence of horizontal movement on the faults. An old erosion surface has been arched and faulted, and certain streams are regarded as antecedent to the vertical uplifts.

The Bruin fault is a major high-angle thrust that dips to the northwest (D. J. Miller, 1959). The Chugach–St. Elias fault is also a thrust which dips northward 30° to 60° and is estimated to have a throw of 10,000 feet. The 1958 movement on the Fairweather fault at the head of Lituya Bay produced scarps in which the vertical displacement was 3½ feet, and the horizontal movement 21½ feet. The southwestern downthrown block moved to the northwest in the horizontal movement (D. J. Miller, 1959).

In general the trenches suggest tension, but their straightness or smoothly curved nature suggests, if not demands, the interpretation of major horizontal displacement. The Farewell–Shakwak fault seems to have cut the Alaska Range and displaced the south side westward (Mt. McKinley block) and the north side eastward, leaving the opposite facing escarpments in nice alignment. These topographic features remind one of features of the San Andreas fault.

The pronounced and fresh-looking topographic features can only mean that major movements on the faults have occurred in Late Cenozoic time. If the faults are like the San Andreas system, then we might suspect a beginning of displacement as far back as the Cretaceous.

The Farewell fault has been connected with the Shakwak trench fault

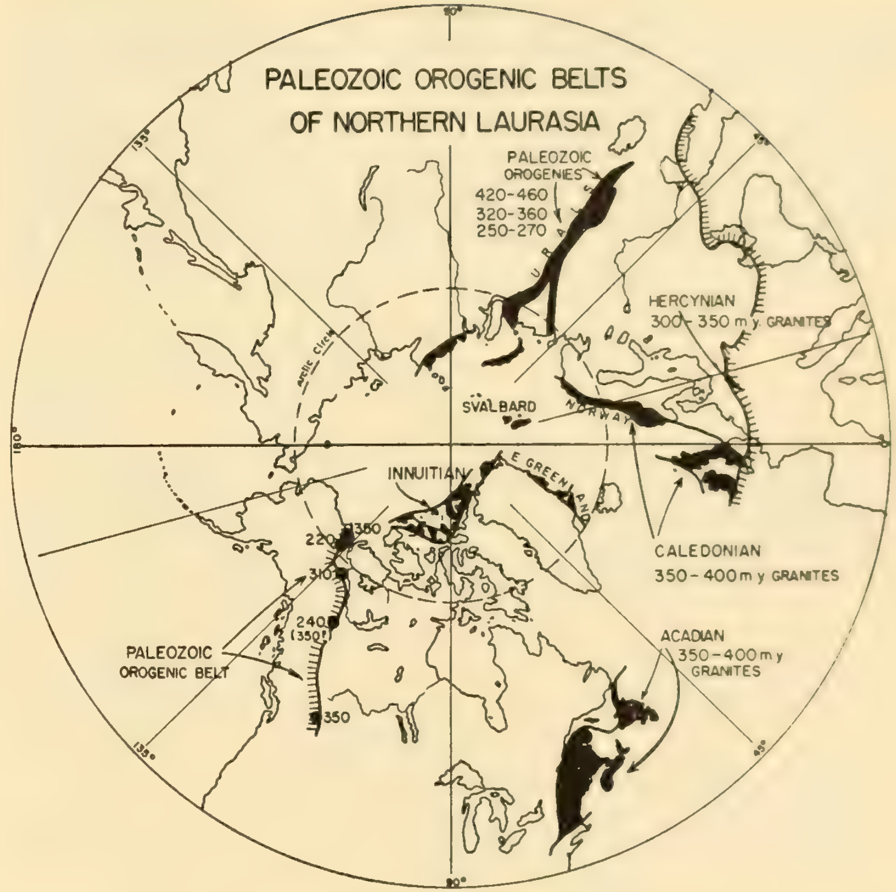


Fig. 39.17. Absolute dates from granites in the orogenic belts around the Arctic. Map kindly supplied by R. E. Folinsbee. See Baadsgaard, Folinsbee and Lipson, 1960.

zone by St. Amand (1957) and the entire system, including the projection down the Lynn Canal, called the Denali fault, for Mt. Denali, the Indian name for Mt. McKinley. He extended the fault down the Alexander Archipelago, along the outer side of the Queen Charlotte Islands, and connected it with the San Andreas fault off Cape Mendocino. Pursuant to the thesis that the Pacific block is in counterclockwise rotation, as

presented by Benioff and depicted in Chapter 32, the Denali fault is presumed to be of right-lateral strike-slip movement, and to have translated rocks 150 miles along the fault.

The principal evidence for right-lateral slip, other than the possibility of the existence of a great fault with smooth arcuate curvature is that of first motion seismograms, and these, according to St. Amand, "indicate that the north Pacific Basin, from Baja California to the Kurile Islands at least, is and has been for a long time, rotating counter-clockwise."

Not much can be said to resolve the problem. St. Amand's interpretation may serve to evoke careful observation of the fault features by those who work in the field along them, and eventually it may be said either that horizontal motion has or has not been appreciable. It may be noted that the other great fault zones of the system were not recognized by St. Amand.

40.

CANADIAN ARCTIC

GEOGRAPHY AND GEOLOGIC PROVINCES OF THE ARCTIC ARCHIPELAGO

The Canadian Arctic Archipelago is a vast domain of islands, channels, bays, gulfs, and peninsulas. It is approximately 1500 miles wide and 1500 miles long, and represents the partly and gently submerged northern margin of the North American continent. The major features of its geography and relief may be seen in Fig. 40.1.

It has been divided into four geologic provinces, namely, (1) the northern part of the Canadian shield, (2) the Arctic Lowlands and Plateaus, (3) the Innuitian region (fold belts), and (4) the Arctic Coastal Plain.

The provinces can be better understood if the entire region is assumed to be emergent because they represent the grouping of the islands or parts of the islands into belts or regions of common geology. See Fig. 40.2.

The sedimentary provinces (eugeosyncline, miogeosyncline, and stable interior) have been defined fairly well, and these should be distinguished from and related to the geologic provinces listed above. Also a younger sedimentary basin (epigeosyncline) has been recognized reposing on parts of the older fold belts and geosynclinal divisions. See Fig. 39.13.

LOWLANDS AND PLATEAUS

Definition

The Lowlands and Plateaus province consists of shelf sediments and intercratonic basins, and is the northern counterpart of the Central Stable Region of the continent. It was called the Arctic Stable Region in the first edition of this book. It will here be considered to include the arches of Precambrian rocks that separate some of the basins.

Areas of Precambrian Rock

The exposures of Precambrian rock of the Canadian Arctic have been described by Fortier (1957) as follows:

> Three areas of the Shield, namely, the Baffin–Ellesmere Belt, the Melville–Southampton Belt, and the Boothia Arch, are underlain by Archean rocks and smaller amounts of Proterozoic rocks. The other two areas, the Wellington and Minto Arches, are underlain by Proterozoic formations.
>
> The Baffin–Ellesmere Belt is the largest and most easterly of the Precambrian areas. It occupies the larger part of Baffin and Bylot Islands, the eastern part of Devon Island, and stretches about half-way along the east coast of Ellesmere Island. The belt is composed chiefly of Archean gneisses and granitic rocks. The structures of the gneisses are complex but a northwesterly trend is prevalent in southern Baffin Island. Proterozoic strata are found in northern Baffin Island and are gently flexed along northwesterly to northerly trending axes. Flat-lying or gently inclined Proterozoic strata also occur at the north end of the belt on Ellesmere Island.
>
> The Melville–Southampton Belt underlies almost all of Melville Peninsula and eastern Southampton Island, continues across Fury and Hecla Strait, and obviously connects with the Baffin–Ellesmere Belt. Little is known about the

Fig. 40.1. Provisional physiographic divisions of the Arctic Archipelago. Reproduced from Fortier, 1959.

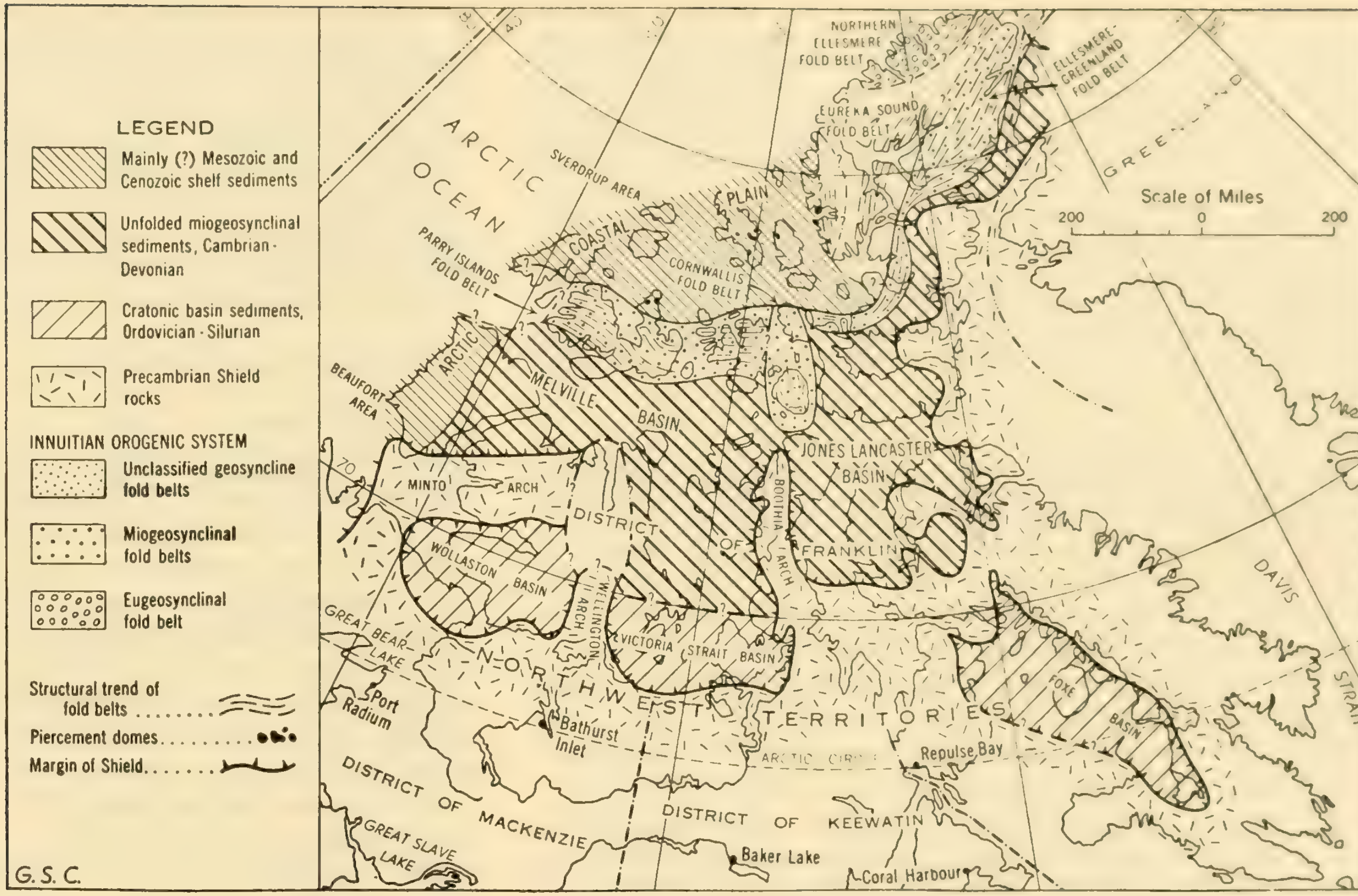


Fig. 40.2. Structural-stratigraphic elements of Arctic Archipelago. Reproduced from Thorsteinsson, 1959, which is revised after Fortier, McNair, and Thorsteinsson, 1954.

geology of this area, but it is apparently underlain mainly by Archean rocks with Protozoic strata along and north of the strait.

The Boothia Arch occupies most of Boothia Peninsula, the western part of Somerset Island, and fringes the southern part of the western shores of Peel Sound. It appears to be mainly of granitic rocks and gneisses, much folded along a northerly to northeasterly regional trend. In the northern part it is flanked apparently by Proterozoic strata which appear to form the outer limbs of a geanticline. This northerly structure may have been in part the effect of a late Silurian orogeny which has affected lower Paleozoic strata adjacent to the Precambrian formations. Basic dykes similar to the so-called diabase dykes so widespread on the mainland Canadian Shield occur throughout the Precambrian of Baffin, Devon, southern Ellesmere, Somerset, and Prescott Islands, and of Boothia Peninsula. They are the youngest Precambrian rocks and their predominant orientation is northwesterly.

The Wellington Arch, in southern Victoria Island, is apparently made exclusively of Proterozoic rocks in obvious extension of the Proterozoic strata of Kent Peninsula and of Bathurst Inlet on the mainland. It trends northerly through Washburn Lake and possibly joins the Minto Arch.

The Minto Arch is much more extensive. It stretches from southern Banks Island across the northern part of Amundsen Gulf to the west coast of Victoria

Island, between the west half of the north shore of Prince Albert Sound and the North shore of Walker Bay. Thence it crosses Victoria Island to its northeastern part, where it probably stretches from Richard Collinson Inlet to the west part of Goldsmith Channel, from which it trends southerly, being possibly within 40 miles of the east coast at Greely Haven. Magnetic data suggest that the Precambrian extends, at shallow depth beneath a thin cover of Paleozoic strata, from the latter locality to the Precambrian of eastern Prince of Wales Island. The rocks of the Minto Arch appear to be entirely Proterozoic and include sedimentary strata in part intercalated with lava and sills. The strata trend northeasterly to northerly and, over most of the belt, form undulatory folds with gentle dips, although in some areas the beds are practically flat lying. In the south half of Wollaston Peninsula unmapped rocks of reddish colour, as observed from aircraft, form many ridges of uniform elevation and oriented east to northeast. Possibly these are Proterozoic formations similar to those of the Minto Arch.

Basins

The basins may be divided into two kinds, those of the miogeosynclinal sedimentary province, and those in the shield (intercratonic). Those considered miogeosynclinal are the Jones–Lancaster and Melville basins, and those of the intercratonic type are the Wollaston, Victoria Strait, and Foxe basins. See Figs. 39.13 and 40.2. In a version of the sedimentary provinces by A. H. McNair these basins are considered mostly intercratonic, with the miogeosyncline being restricted to the fold belts (map supplied writer by McNair).

The Jones–Lancaster and Melville basins are separated by the Boothia arch. According to Fortier (1957):

> They extend from Banks Island to Bache Peninsula, midway along the east coast of Ellesmere Island and accordingly, lie mainly between the outer areas of the Shield and the Innuitian Region. Most of the strata of the Jones–Lancaster Basin Range in age from Cambrian to Devonian but may include rocks of Tertiary age. Although normal faults and a few folds are present, the strata throughout most of the basin dip gently away from the Shield areas and towards the Innuitian Region. Thus, in the northern part of the basin on Ellesmere Island, the regional dip is northerly, farther south it is westerly, and near the south coast of the island it is northwesterly. On Devon Island the dip is westerly and on Brodeur Peninsula of Baffin Island it is northwesterly. Near the Innuitian Region, however, at least some of the beds are flexed into folds which are probably related to the orogenies that affected that region, but are on a smaller scale. Such folds are found, for instance, on Somerset Island. In the northwestern and south central parts of Somerset Island and in the northeastern part of Prince of Wales Island, that is, on each side of the Boothia Arch, a late Silurian or early Devonian conglomerate is made of detritus derived from the Precambrian rocks of the arch. However, the arch is presently separated from the conglomerate by a wide exposure of earlier Paleozoic strata, the gentle flexure of which, at least in the east, may have been contemporaneous with the uplift and denudation of the arch and with the deposition of the conglomerate. Little is known of the Melville Basin, except for the above conglomerate, but Silurian strata are apparently widespread in its eastern part and Devonian strata occur in its western part. North of the Minto Arch, the strata on northwestern Victoria Island regionally dip gently to the northwest; on northern Banks Island they are flexed in gentle, southerly trending folds; and on southwestern Melville Island they are flat lying to gently flexed.

FOLD BELTS—THE INNUITIAN REGION

Nature and Distribution

A belt of strong deformation extends from North Greenland southwesterly through the Arctic Archipelago to the Parry Islands. It consists of folds of mid- and late Paleozoic age (pre-Middle Pennsylvanian) developed in eugeosynclinal and miogeosynclinal strata, and structures of late Mesozoic and Tertiary age in basin beds laid down on the older orogenic complex. The fold belt in the miogeosynclinal strata (Fig. 40.2) is divided into a western segment, the Parry Islands fold belt, and an eastern, the Ellesmere–Greenland fold belt, by a transverse zone of structures, the Cornwallis fold belt. The Cornwallis fold belt is a northern continuation of the Boothia arch.

The Northern Ellesmere Island fold belt is regarded as deformed and metamorphosed eugeosynclinal strata.

The Eureka fold belt is the northeastern part of the Sverdrup basin which is composed of Late Pennsylvanian and younger beds laid down on the deformed eugeosyncline and miogeosyncline.

Parry Islands Fold Belt

The Parry Islands Fold Belt includes, in its eastern part, at least 1,800 feet of calcareous and dolomite mudstone and shale, in part silty, overlain by 3,000 feet of further Silurian graptolitic, argillaceous and calcareous, fine-grain sandstone. These are conformably overlain by 1,200 feet of Silurian or Lower Devonian calcareous and argillaceous sandstone, 800 feet of Lower Devonian shale and these are followed by a Middle and Upper Devonian sequence similar

to that found on southern Ellesmere Island. The shaly equivalent of the graptolitic rocks might occur in the unexplored southern part of Bathurst Island, as some are known on southern Cornwallis and northern Prince of Wales Islands. The western part of the fold belt includes over 1,000 feet of Ordovician and possibly earlier limestone and conglomerate. In part the Ordovician, Cornwallis formation, with over 1,500 feet of shaly limestone and dolomite, is overlain by 2,500 feet of graptolite shale of the Cape Phillips formation; in other parts are Ordovician and Silurian black graptolitic shale, argillite, chert with minor dolomite, in all some 3000 feet thick; still elsewhere are over 6,000 feet of Silurian and possibly Ordovician dolomite and limestone. The Devonian includes up to 8,000 feet of marine and non-marine sandstone, siltstone, and shale; 2,500 feet of non-marine sandstone, and 4,000 feet of non-marine sandstone, shale, and coal, with marine bands.

The Parry Islands Fold Belt was folded before the deposition of the Pennsylvanian. The synclines generally have broad troughs and the anticlines have narrow crests with the more steeply dipping strata close to the crests. Many folds are doubly plunging but closures are still to be determined. Where the belt abuts the transversal Cornwallis Fold Belt, deformation has resulted in folds of various shapes and orientations, some folds are almost circular in shape, others have curving axes, and some are aligned parallel to those of the Cornwallis belt (Fortier, 1959).

Ellesmere–Greenland Fold Belt

The Ellesmere–Greenland Fold Belt comprises at least 870 feet of Middle Cambrian limestone and minor shale; 4,800 feet of limestone and impure limestone with gypsiferous beds, possibly ranging from Cambrian to Middle Ordovician; up to 4,400 feet of the Middle Ordovician Corwallis formation; 3,700 feet of the Ordovician to Middle Silurian Allen Bay formation; at the very least 1,500 feet of Middle to Upper Silurian limestone, silty limestone, and dolomite. The Ordovician to Upper Silurian graptolitic Cape Phillips formation, at least 2,300 feet thick, has been located only north of Baumann Fiord and approximately halfway across the fold belt. Either in the Upper Silurian and/or the Lower Devonian are numerous sections correlated with difficulty either because of the nature of their fauna or their unfossiliferous nature. They differentially contain dolomite, sandstone, limestone, siltstone and shale in various degrees of purity, and vary in thickness, the thicker section measuring some 4,000 feet. Marine calcareous shale and siltstone, over 1,000 feet thick, are probably Lower Devonian. The Middle Devonian includes up to 3,800 feet of limestone, dolomite, and calcareous shale, in part with coral biostromes and bioherms, overlain by a maximum of 2,900 feet of marine limestone, sandy limestone, sandy shale, and sandstone. The Upper Devonian over 10,000 feet thick, is largely made of non-marine sandstone and shale with thin seams of bituminous coal.

The above formations were folded, in the southern and western parts of the Ellesmere–Greenland Fold Belt, prior to the Pennsylvanian, but in the eastern part of the belt they were folded only in the Tertiary, conformably with non-marine Tertiary and possibly Upper Cretaceous sandstone, shale, and coal. In general the lower Paleozoic miogeosyncline is the most deformed in this belt, folds are symmetrical and asymmetrical, some are overturned, thrust faults and high angle faults are known. The deformation has been more severe northward, where the stratigraphy is less known and some metamorphism produced slates, phyllites, and fine-grained schists (Fortier, 1959).

Northern Ellesmere Fold Belt

The Northern Ellesmere Fold Belt underlies the northern coastal area of Ellesmere Island and apparently extends to northwestern Axel Heiberg Island. The rocks comprise sedimentary and volcanic material possibly ranging from the Precambrian to the Tertiary. A part of the belt includes gneisses and intrusions that vary from granitic to ultrabasic. These are undated but it is probable that they are Precambrian in age and were deformed during that time. Some volcanic rocks are pre-Permo-Carboniferous, either Silurian or Devonian, and are adjoined by greywackes. These rocks and Ordovician beds are mildly metamorphosed but have complex structures that probably resulted from the Variscan orogeny. Widespread outliers of mildly folded Permo-Carboniferous strata unconformably overlie older rocks of more complex structures and indicate that the Late Mesozoic and Tertiary deformation extended to those parts (Fortier, 1957).

After Fortier wrote the above paragraph, a note was published by Blackadar (1960) on a group of gneisses and migmatites between Cape Aldrich and Markham Inlet which he had named the Cape Columbia group. These had been demonstrated on stratigraphic grounds to be older than Middle Ordovician. A potassium-argon analysis was made on a biotite-rich gneiss and an age of 545 m.y. was obtained. This is latest Precambrian or earliest Cambrian. Blackadar concludes that the orogeny formed a landmass from which clastic sediments in the Parry Islands and Ellesmere Island came. By the close of the Paleozoic era the Cape Columbia terrane had been lowered and Permian limestones were deposited on it.

Cornwallis Fold Belt

McNair (1960) has reported as follows on the Cornwallis Fold Belt (see Fig. 40.3):

Two sets of regional structures meet on eastern Bathurst Island. The older, north-south trending Cornwallis folds are characterized by vertical movement

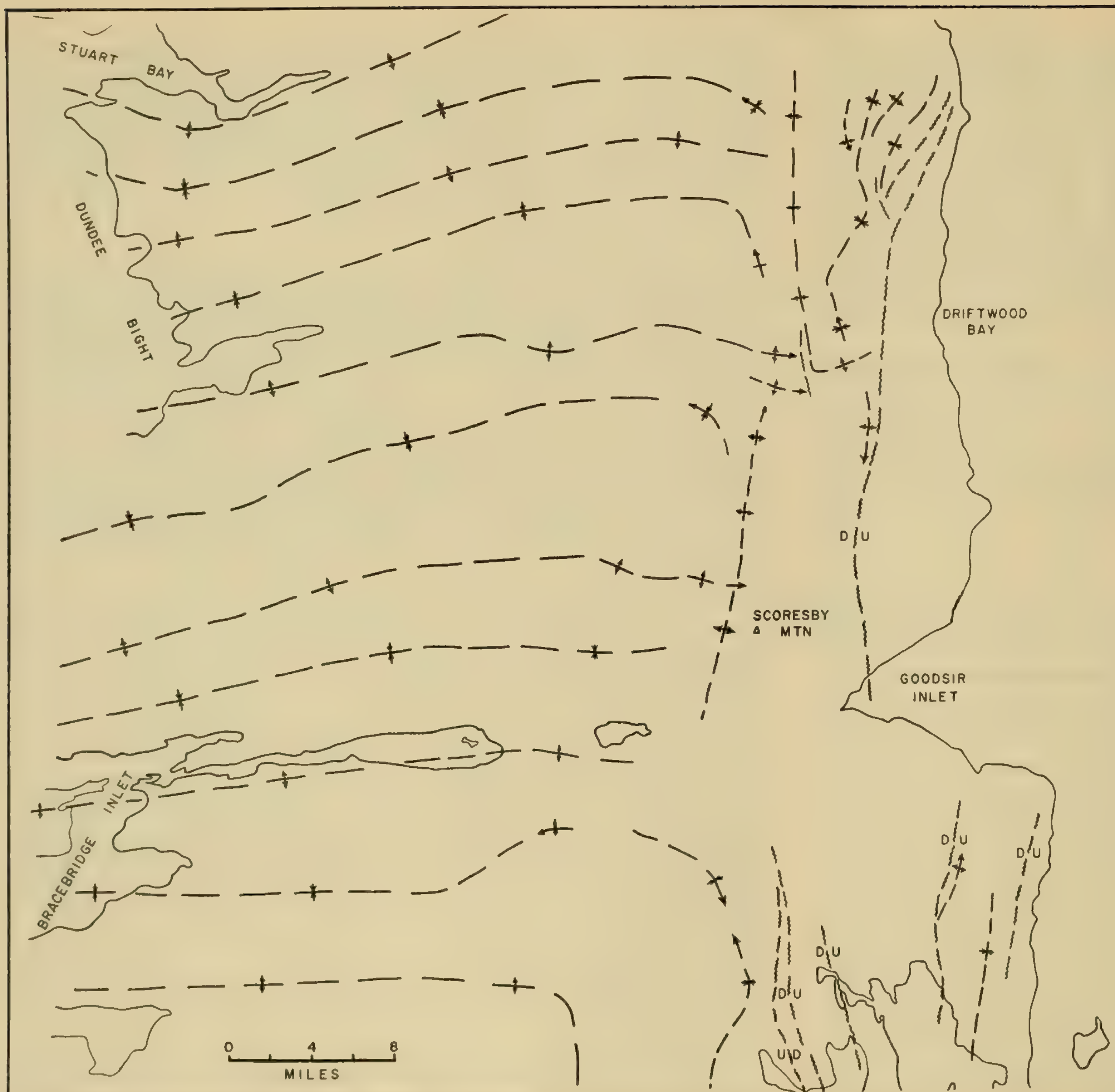


Fig. 40.3. Fold axes and faults of east-central Bathurst Island. Kindly supplied by A. D. McNair, Dartmouth College.

and appear to be of cratonic origin. These range from narrow, steep-flanked anticlines to broad synclines and anticlines. Conglomerates and two unconformities indicates that the Cornwallis folds had an initial, intermittent development extending from Middle Silurian to the Middle Devonian. During short times of stability in the Silurian many small reefs grew along the margins of some anticlines.

The east-west Parry Islands miogeosynclinal fold belt consists of long parallel folds which decrease in amplitude eastward toward the Cornwallis folds. However, at several places the Cornwallis structures are relatively highly deformed by the east-west folds. The Parry Islands belt was deformed in the latest Devonian or during the Mississippian.

The final phase of deformation occurred as persistent north-south postorogenic faults. In the southeastern part of Bathurst Island the faults controlled the emplacement of small sills, dykes and plugs of olivine basalt.

Sverdrup Basin

The Sverdrup Basin includes a voluminous sequence of Pennsylvanian to Tertiary beds which have been mainly deformed in Tertiary time. In the best exposed and apparently thicker part of the basin, the units of the sequence, although varying in thickness, appear essentially conformable. At the periphery of the basin there are unconformities, disconformities, oversteps and some facies developments.

In the southern and eastern peripheries of the basin, the Permo-Carboniferous commonly includes units of limestone, units of sandstone with layers of conglomerates, and lesser units of shale. In the northern part of Axel Heiberg Island, the Permo-Carboniferous includes volcanic measures. Nearby, Permian limestone is at least 5,000 feet thick. Across the middle part of the island, the basin contains the following units: Permian siltstone with lesser shale and silty shale, 4,000 feet thick; Middle, Upper, and probable Lower Triassic shale with siltstone and sandstone, 10,000 feet thick; Upper Triassic, marine, and possibly Lower Jurassic, non-marine sandstone, shale, siltstone, with carbonaceous film in the upper part, up to 5,600 feet thick; Jurassic marine shale up to 900 feet thick, non-marine sandstone and lesser marine shale up to 1,300 feet thick; Jurassic and Cretaceous shale as thick as 2,500 feet; Lower Cretaceous sandstone with a maximum thickness of 4,500 locally with a 200-foot stratum of volcanic breccia, and shale in thicknesses reaching 3,000 feet; Lower or Upper Cretaceous sandstone and shale, over 700 feet thick, locally overlain by basalt flows up to 600 feet thick; Upper Cretaceous shale, as thick as 1200 feet and conformably overlain by Tertiary and possibly Cretaceous non-marine siltstone, sandstone and silty shale with coal, over 8,000 feet thick. In the Ringnes and Cornwallis Islands these or similar units down to the Upper Triassic occur but in somewhat lesser thicknesses. Facies indicate an eastern and southern source for most of the Mesozoic sediments. Gabbro sills and lesser dykes are common in some units and are most numerous in the region of Eureka Sound. They are as far widespread as from Melville Island to the region of Baumann Fiord, from Ellef Ringnes to the northeast coast of Ellesmere Island. There is no regional metamorphism and any alteration related to these intrusions is limited to a few feet in the country rocks.

The strata of the basin have been folded in Tertiary time. From the Ringnes Island eastward the deformation has been more marked, and the northwesterly and northerly trending folds form the Eureka Sound Fold Belt. On the southern part of Axel Heiberg Island many folds are of the "box" type. The regional plunge of the folds is inward to the basin, that is towards the longitudinal axis of the basin. This axis on the Ringnes and Axel Heiberg Islands is generally the locus of the youngest formations. Coinciding with this axis from northernmost Melville Island, across the Ringnes Island, middle Axel Heiberg Island, the eastern part of this island along Eureka Sound to Nansen Sound and Ellesmere Island is a zone of diapiric intrusions of gypsum of Pennsylvanian and/or Permian age. Most of these intrusions are in the crestal area of the Tertiary folds. There is a rough parallelism between the longitudinal axis of the Sverdrup Basin, the zone of diapiric intrusions, and the trend of the deformed lower Paleozoic miogeosyncline. It would thus appear that the Paleozoic orogeny had long range effects in that not only was it a factor in the formation of the depression in which Permo-Carboniferous evaporites were eventually laid down, but also it ultimately had some bearing on Tertiary tectonism (Fortier, 1959).

See the summary by Tozer (1960).

ARCTIC COASTAL PLAIN

The Arctic Coastal Plain covers the western part of Banks Island, the western and northwestern parts of Prince Patrick Island and probably extends to the northwestern parts of Brock, Borden, and Ellef Ringnes Islands. The rocks include Cenozoic beds unconformably covering Mesozoic strata and, south of the Sverdrup Basin, comprise Cretaceous and possibly Jurassic formations apparently overlapping Devonian strata (Fortier, 1957).

CORRELATION WITH ALASKA AND THE YUKON

Reference to Fig. 39.13 will bring to one's attention the following possible correlations of the geologic provinces of the Arctic Archipelago and Alaska and the Yukon. The Pennsylvanian and Permian of the Sverdrup basin would appear to have a tie with the Late Pennsylvanian and Permian of the basins of northeastern Alaska and northern Yukon. The closeness of the shelf margin to the present shore leaves little room, however, to connect them into a continuous basin. The unconformity

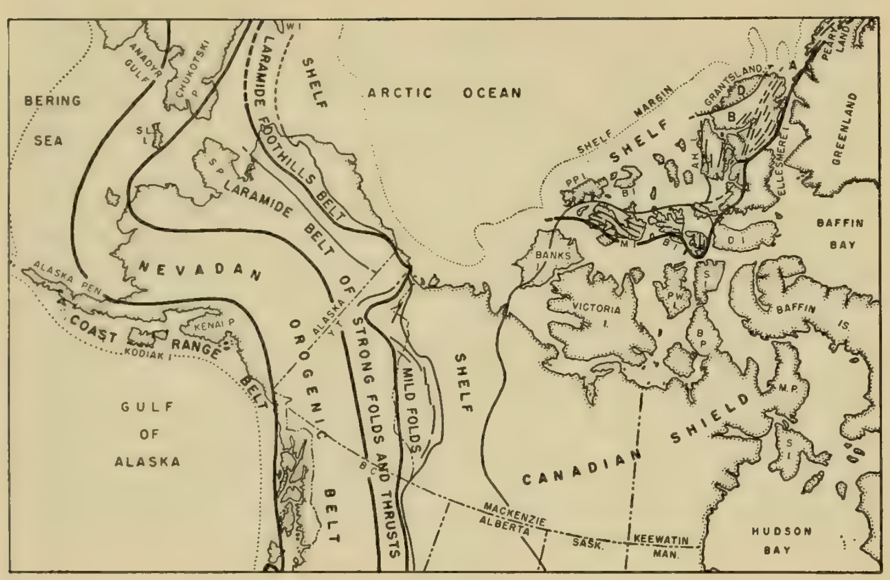


Fig. 40.4. Belts of deformation of northern North America. The fold belts of the Arctic Archipelago are after Fortier *et al.*, 1954. A, Ellesmere-Greenland fold belt; B, Eureka Sound fold belt; C, Parry Island fold belt; C, northern Ellesmere fold belt (in eugeosynclinal sediments); D, Coastal Plain. For details of the Cordilleran belts of deformation see Fig. 39.2.

below the Late Pennsylvanian strata and the older formations is common in both regions and draws them together in a common province.

The northern Ellesmere fold belt in rocks of eugeosynclinal character, parts of which are probably Precambrian, may relate to the Paleozoic positive area of northern Alaska. The latter's rocks are only known in well cores and are identified as argillite, probably Precambrian in age. Again, the continental shelf is fairly narrow from Ellesmere Island to Alaska, and not much room is available under it to connect the Precambrian terranes. Certain authors have presumed the lands to have extended northward into areas now of deep water, and imagined subsidence in the order of 10,000 feet to have occurred, but as we shall see, this is probably not possible.

The Laramide structures of Alaska extend to the Arctic shoreline in northeastern Alaska and northern Yukon, as if perhaps, they once continued northeastward under the continental shelf. See Fig. 40.4. Structures of the same age in the Eureka Sound Fold Belt suggest that the two may have been continuous. There seems no way, however, to prove or demonstrate this postulate.

The subject of possible connections will be pursued farther on following pages when the origin of the Arctic Basin is considered.

PLEISTOCENE EPEIROGENY AND CLIMATIC CHANGES

Washburn (1947) reports that Victoria Island has emerged at least 500 feet since the last glaciation, as demonstrated by raised strand lines and marine fossils. In addition he believes the whole of the Arctic Archipelago has suffered comparable movements. Elevated beaches are reported by G. M. Stanley (personal communication) up to 900 feet above sea level along the east coast by Hudson Bay.

Continental ice sheets formerly covered all Arctic Canada east of the Cordillera except some of the western Queen Elizabeth Islands (Craig and Fyles, 1960). The elevated strand lines represents an isostatic adjustment following the melting of the ice, and such emergence was almost complete before the final eustatic rise of the sea.

Numerous Tertiary deposits have been found in the Arctic region, and fortunately most of them carry coal beds and fossil plants. By reconstructing the character and distribution of the Tertiary flora from the Arctic to the temperate regions of the northern hemisphere, with particular reference to the redwood Sequoia, Chaney (1940) concluded as follows (Fig. 40.5); the Arctic cooled gradually from late Eocene to the Pleistocene with a slight reversal in mid-Miocene (personal communication, E. Dorf), and the temperate rain forests shifted southward across 60° of latitude incident to the cooling. He postulates that the gradual cooling was caused by and was coincident with a gradual uplift of the continent.

OROGENIC BELTS OF GREENLAND

Paleozoic

East Greenland north of 70° N. Lat. is marked by a belt of Caledonian (Late Silurian and Early Devonian) orogeny, and another belt of orogeny

of the same age extends across the northern margin of the great island. The East Greenland fold belt developed during three phases (Koch, 1961):

1. Orogeny of Silurian (?) age affected the entire east coast, with thrusting toward the west and extensive granitization.

2. Deformation south of 76° N. Lat., in places closely related to intrusive granite bodies, occurred in Devonian time.

3. Subsiding basins were filled with thick deposits of molasses-type sediments in the Middle and Late Devonian, in the Carboniferous and in the Early Permian. They attest times of nearby crustal unrest and elevation, but the Devonian detritus was mildly deformed itself in two episodes, one in Early Carboniferous and one in Early Permian. The entire east coastal area was strongly affected by faulting, especially during the Carboniferous.

Mesozoic and Tertiary

A marine transgression in Late Permian time covered large areas along the coast, and this was followed by several Mesozoic transgressions. Many of the old faults were reactivated in the Tertiary.

A large basalt field of Late Cretaceous and early Tertiary age occurs in the east-central part of Greenland (Fig. 40.5) and of this region Wager (1947) writes:

> Subsequent to the forging of the metamorphic complex which probably took place in Pre-Cambrian times, the area was for long dominantly subjected to upward movement with concomitant erosion. Towards the end of the Mesozoic era, when next there is definite information, the area seems to have been of subdued relief and near sea level. In the Kangerdlugssuaq area a local marine transgression of approximately Senonian age produced thin sediments resting on the metamorphic complex, and a similar and perhaps contemporaneous marine transgression took place further south on what is now Kap Gustav Holm. Within a short time of the maximum development of the Cretaceous transgression volcanic activity broke out in the Kangerdlugssuaq region giving the Lower Lavas and Tuffs.
>
> The Lower Lavas and Tuffs of latest Cretaceous or very early Eocene age, mark the beginning of intensive igneous activity in East Greenland, extending in a N.N.E. direction over a distance of 1,200 km., from 66° to 75° N. Southwards, the coast line has the same N.N.E. direction and there are many basic dikes, which almost certainly form part of the same igneous episode.
>
> The eruption of vast quantities of basalt to give the Plateau Basalt Series, attaining in places a thickness of certainly 6½ km. and probably a good deal more, is the greatest igneous event in the region judging by the quantity of magma involved. The time taken for the accumulation of the Plateau Basalt Series can be estimated from the fossils found immediately below and above the series as approximately equal to the duration of the Lower Eocene, and this

Fig. 40.5. Upper map, very generalized distribution of seas and lands of the Arctic during Triassic and Jurassic times. The seas at any one time were not as extensive as the total distribution shown. Lower map, early Tertiary deposits of the Arctic. The dotted lines are isoflors after Chaney, 1940, and the crosses denote Chaney's Eocene and Oligocene localities, plus a few other localities where "Arctic Miocene" coal beds are known. The ruled area denotes the Greenland-Iceland-Scotland basalt field of early Tertiary time.

may be taken to be of the order of 5–10 million years. The fact that the sediments immediately underlying and overlying the thick Plateau Basalt Series are both of shallow water marine origin shows that during or soon after the extrusion of the basalts there must have been sinking of the basalt pile comparable in amount with its thickness.

Some basic intrusions, e.g., the Skaergaard and Kap Edvard Holm complexes were formed during or soon after the main period of basalt outpouring. This also seems to have been the chief period of sill intrusion although this phase never reached large proportions.

The chief tectonic event affecting the area, namely, the elevation of what is now the coastal mountain belt of East Greenland and the sinking of the area which is now the Denmark Strait, took place subsequently to the formation of the main plateau basalts. The junction between the two areas of differential epeirogenic movement is marked in Middle East Greenland by a flexure of the crust. Where the flexure is intense with dips of more than 10°, a dike swarm is developed which follows the convex part of the flexure. The intensive flexuring and associated dike swarm occur along much of the Middle East Greenland coast and, as it is likely that all the flexuring took place during the same limited period of time, we are provided with a useful method of dating certain local events. The coastal flexure and dike swarm almost certainly came after the formation of the Kap Dalton sediments, which are Middle or Lower Eocene. The main part of the inland doming of Knud Rasmussens Land is considered to have been incidental to the general epeirogenic uplift and to have developed at that time.

Not all of this impressive differential vertical movement is to be ascribed to the coastal flexure stage and it is suggested that the total movement as now determined by the lie of the rocks can be analysed into the following parts:

1. Early slight flexuring due to differential sinking of the lava pile as it accumulated.
2. The main epeirogenic movement and associated flexuring, with a dike swarm where flexuring was sufficiently intense.
3. Possible later up-warping of the edge of the uplifted area as a result of isostatic adjustments to erosion and to the development of the ice cap.

Faults have been recognized on both the west and east coasts of Greenland. The Cape York district of northwest Greenland is especially broken by high-angle faults (Koch, 1929), and the fiords of the west coast about Disko and Umnak bays generally take their courses parallel to faults (Hobbs, 1932). It is not clear, however, that these faults are to be associated with Tertiary land movements. Koch (1935) believes that strong faulting in Tertiary time may be recognized in many places along the eastern coast, and that it is associated with the great volcanic activity just described. The faults have tilted a plane to the west on Molne Land, and may be seen cutting the sediments there. Along the east side of Hurry Inlet are Tertiary faults, and Liverpool Land was doubtless strongly raised in Tertiary time. Although the basalts with their great flexure are not present north and south of the middle east Greenland area, the topography along the coast in the absence of the basalts suggest comparable crustal movement (Wager, 1947). The volcanics of east Greenland, as a number of writers have proposed, must be continuous with the basalt fields of Iceland, the Faeroes, and Scotland; but Wager does not believe that they extend under the ice of Greenland and connect with the basalts of the west coast.

Precambrian

An outline of the Precambrian rocks and history of east Greenland is given below. It is after Koch (1961).

East-Central Greenland

Eleonore Bay group (Proterozoic)	Upper
	Tillite and varved strata, 200–1000 m
	Dolomite and ls, 1100 m
	Psammite, pelite, 3000 m
	Lower
	Tillite, ls, phyllite, 2600–7400 m.

Archean basement

Northeast Greenland

Proterozoic
- Hagens Fiord group, derived from Carolinidian belt
  - Faulting and eruption of basalts
  - Folding and magmatic activity (Carolinidian orogenic belt)
  - Basalt dikes and sills
- Thule group (psammites), 3000 m
  - Greenlandian (semipellites), 3000 + m

Archean basement

ARCTIC OCEAN BASIN

Surrounding Shelves

The floor of the Arctic Ocean is about half shelf and half deep basin. See Fig. 40.6. Off Alaska and the Canadian Arctic Archipelago the shelf

Fig. 40.6. Bathymetric chart of Arctic Ocean. Compiled from Soviet sources as of 1956 by Chief Cartographer, Surveys and Mapping Branch, Dept. of Mines and Technical Surveys for Defense Research Board of Canada, Ottawa, 1957.

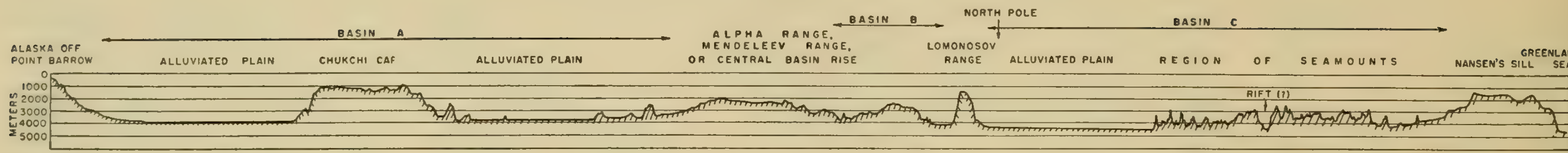


Fig. 40.7. Bathymetric profile across Arctic basin, taken in August, 1958, by SSN (571) *Nautilus*. Drafted from chart kindly supplied by Dietz and Shumway. Basin A is called Beaufort Sea basin by Soviets, and Canada basin by Dietz and Shumway. Basin B is called Makarov basin by Soviets and central Arctic basin by Dietz and Shumway. Basin C is called Nansen basin by Soviets and Eurasia basin by Dietz and Shumway. Nansen's Still is called Nansen Ridge by Dietz and Shumway.

is narrow, but off Eurasia it is very broad. Spitzbergen (Svalbard), Franz Josef Land (Zemlya Frantsa Iosifa), North Land (Severnaya Zemlya), Novaya Zemlya, New Siberian Islands (Novosibirskiye Ostrova), and Wrangel Island (Ostrov Vrangelya) all rise from the shallow but broad shelf north of Siberia and Norway.

Spitzbergen was formerly believed to be tied to northern Greenland by the Nansen sill, but recent soundings show that the sill is broken by a transverse trench with a floor 3100 to 3900 meters deep and about 200 kilometers wide (Hope, 1959b).

Deep Basin

The deep basin is approximately triangular in shape with the base about 1150 miles across and the side from Spitzbergen to Alaska about 1650 miles long. On the basis of post-war soundings, principally by the Russians, the large basin is known to be divided by the Lomonosov Range (or Ridge) which extends from the New Siberian Islands to Greenland and Ellesmere Island, a distance of 1800 kilometers. Its peaks rise 2500 to 3000 meters above the adjacent ocean floor, and the highest peak yet sounded is 954 meters below the ocean surface. Saddles to a depth of 1500 meters, spurs, and steep slopes are characteristic.

On the Alaskan side of the Lomonosov Range another range was discovered by the United States drifting ice station Alpha. It has provisionally been called the Alpha Range by Hope (1959a). Its extent is not known and its relief appears to be less than the Lomonosov Range. Its apparent plateau-like top rises to 2300 meters below sea level. The two ranges then divide the major deep basin into three sub-basins which have not yet been named officially. They will be referred to here as basins A, B, and C. The scientists of the U.S.S.R. and of the United States respectively have called them as follows; Basin A, Beaufort Sea Basin and Canada Basin; Basin B, Makarov Basin and Central Arctic Basin; and Basin C, Nansen Basin and Eurasia Basin (personal communication V. N. Sachs and charts prepared by Dietz and Shumway). The Alpha Range is called the Mendeleev Range by the Russian scientists, and on unpublished charts by Dietz and Shumway, the Central Basin Rise.

Basin C, which lies north of the Greenland, Barents, Kara, and Laptev seas, is the deepest of the three and has a maximum depth of over 5220 meters. Basin B on the opposite side of the Lomonosov Range, has depths over 4000 meters. Basin A which lies north of the Chukchi and Beaufort seas has depths up to 3820 meters.

The sonic depth profile recorded by the submarine *Nautilus* across the Arctic Ocean, is summarized in Fig. 40.7. It extends from a point north of Point Barrow directly to the North Pole and beyond to the middle of Basin C, and thence southwesterly to Nansen's Sill between Spitzbergen and Greenland. Its features should be noted, and in succeeding paragraphs they will be referred to.

Seismic studies over the Alpha Range by Hunkins (1961) indicate the boundary between the 4.7 km/sec layer and the basaltic layer at about 5 kilometers which is less than that shown by Demenitsckya. Relief of the rise is rugged and apparently the result of block faulting. The constitution of the crust is similar to that of the Atlantic Ocean floor.

Nature of Crust under Deep Basin

A seismic surface wave of unusually large amplitude (Lg wave) was recognized by Press and Ewing in 1952, and it was noted to have the characteristics of traveling only over paths of continental structure. It does not propagate across oceanic crust. Oliver *et al.* (1955) subsequently studied the Lg wave paths across the Arctic region and concluded that the Arctic basin was floored by ocean crust, and that it could not be sunken continental crust as had been postulated by Soviet geologists and Eardley. The subsidence theory will be considered later.

Figure 40.8 is an interpretation of the crustal constitution across the Arctic basin from Franz Josef Land to Alaska by Demenitsckya (1958). A thickening of the basaltic layer under both the Lomonosov and Alpha ranges is conspicuous, as well as the existence of a 5-kilometer thick "granitic" lens under each. The crust under the basins is typically oceanic.

Theories of the Origin of the Arctic Basin

Permanency of the Basin. In about 1860 James D. Dana began to teach that the continents and ocean basins are permanent features of the earth's crust. He contended that in the main the ocean basins have been sinking and the continents rising, but several continental fragments have subsided. Fifty years later Charles Schuchert (1916) in his studies of paleogeography was foremost in supporting Dana. He said:

> Now, however, geologists are holding more and more to the hypothesis that the earth periodically shrinks, and each time it does so some parts or all of the continents rise more or less; but that in the main there is subsidence of the ocean bottoms equal in amount to the rising land-masses, that the water of the hydrosphere is constantly increasing in amount, and that even though the continents are in the main permanent, yet they are partially breaking down into the oceanic basins.

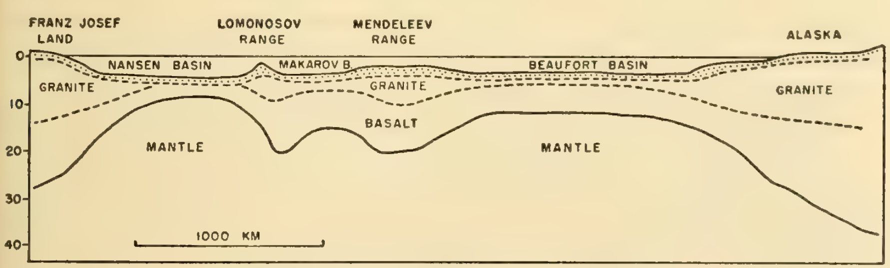


Fig. 40.8. Crustal structure of the Arctic basin, after Demenitsckya, 1958. Basin names and Mendeleev Range name, have been added according to information from V. N. Sachs personal communication. Mendeleev Range has been called the Alpha Range by Americans.

Reference is made by Schuchert to the permanency of the North Atlantic Ocean basin, but we can only presume that he considered the Arctic Ocean basin a permanent feature; his maps are not definitive about the Arctic.

Subsidence Theory. In the period of 1930 to 1950 and beyond, the Russians, beginning with Shatski (Hope, 1959a) considered the Arctic Ocean basin as a sunken region, once emergent. The sunken crust was called the Hyperborean shield, and the depression as the Hyperborean basin. Later, associates of Shatski referred to the shield as a massif or platform. The sunken platform was postulated as a result of an envisioned belt of Mesozoic folding encircling the basin, as a once resistant shield.

A development of the thesis that the fold belts which extend to the Arctic Ocean cross the shelves and deep basin is shown in Fig. 40.9. Here, Sachs *et al.* (Hope, 1959a) show Caledonian, Hercynian, and Alpine fold belts extending across the deep basin, particularly where the Lomonosov Range and Nansen Basin (C) now exist. They postulate, of necessity, that the fold belts have sunken to form the deep basins.

In 1948 Eardley reviewed the geology of the lands around the Arctic Ocean basin and concurred with the Russians that the basin was a sunken region which in Precambrian and perhaps early Paleozoic times had been land. The broad shelves and relatively small size of the basin, the facing Precambrian shields (Canadian, Greenland, Russian–Baltic, and Angara), the Paleozoic orogenic belts that project to and under (?) the Arctic Ocean (Ural–Nova Semlya, Norway and Spitzbergen, East Greenland, Canadian Arctic Archipelago, Northland, and New Siberian) suggested to him that the region was once land and beginning in Paleozoic time has foundered. Paleozoic geosynclinal conditions in Alaska seemed to require a source for some of the sediments north of land today, where water is fairly deep. Paleozoic fossil faunas common to North America

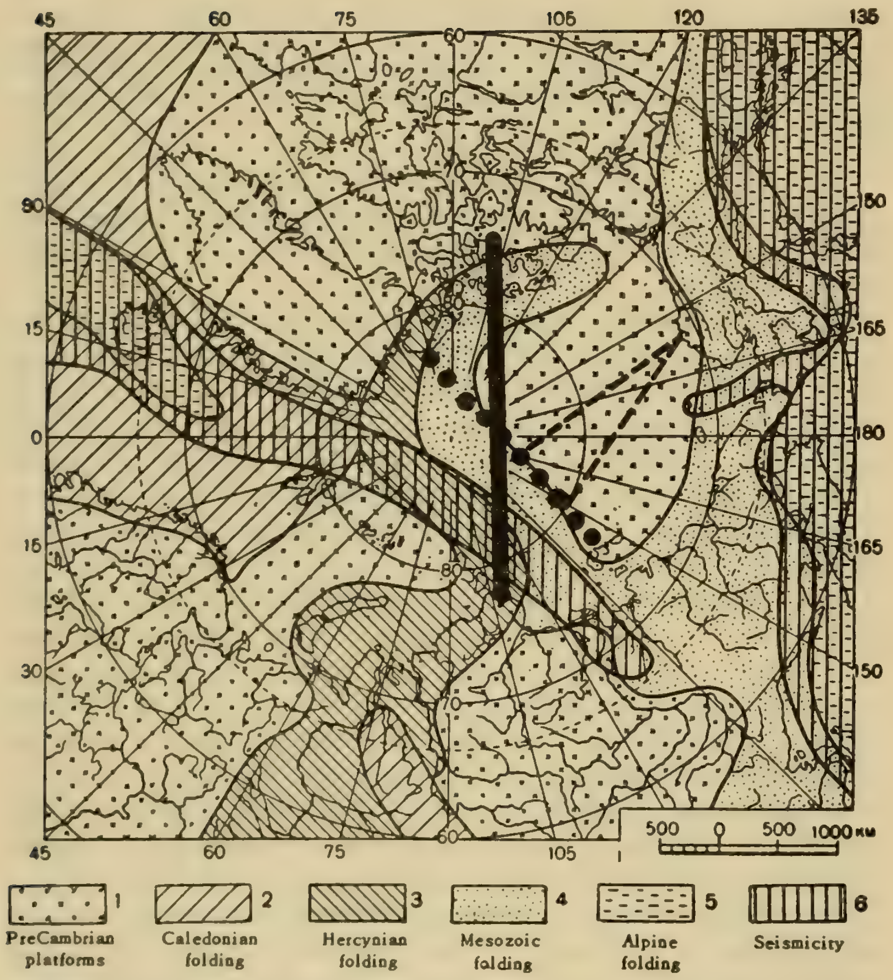


Fig. 40.9. Postulate of fold belts across the Arctic basin, by Sachs, Belov, and Lapina (1955). Reviewed in English by Hope (1959a). The broad black line is the great Arctic magnetic anomaly, and the row of dots the Lomonosov Range.

and Eurasia find an explanation in the possible shallow sea migration routes bordering the once emergent and later subsiding regions.

Continental Drift. Although two or three geologists before him had suggested without much documentation the concept of horizontal shifting of continental fragments Taylor, in 1910, is generally given credit for "specifically advocating continental drift" (Van Waterschoot van der Gracht, 1928). Wegener first addressed the subject in 1912 but it was not until his comprehensive study, *Die Enstehung der Kontinente und Ozeane,* was published in 1922, that the theory became of international concern. Although many European geologists supported the concept of continental drift in one form or another, most American geologists continued to favor the Dana–Schuchert concept of permanency of the continents and ocean basins. The theories of continental drift, however, focused attention on the Arctic Ocean basin, and Taylor in particular dwelt specifically on it. He postulated drift toward the equator and away from the North Pole. Figure 40.10 illustrates the general concept and his view of the origin of the Arctic Ocean depression as a "disjunctive basin." Eurasia and North America were once together over the North Polar region as one great continent, but pulled apart leaving the Arctic basin as one of the disjunctive depressions. Greenland was considered a fragment left between the Baffin Sea basin and the Greenland Sea depression as Europe drifted away from North America. The approximate extent of the continental shelves and the deep basin in the Arctic had been established by Nansen and other explorers but no detail of the bottom topography was known at this time.

Wegener gave more attention to the southern hemisphere and Antarctica than to the Arctic Ocean basin, and we are left to examine his maps to discern his thoughts about the origin of the Arctic basin. The maps show an existing ocean there, although small, before the breakup occurred. In a major publication in 1924, however, Köppen and Wegener show the small basin to enlarge appreciably as North America, hinging in the North Polar region drifted westward and away from Europe.

By the time of Wegener's major publications the concept of a layered crust had become established. The continents were made up of a silicic and lighter upper layer, the sial, resting on a mafic and heavier layer, the sima, and when a continent broke and its parts drifted away from each other, it was the sial that parted and drifted over the sima, leaving a crust made up only of the sima. This, for isostatic reasons, was also a basin. Hence, according to Köppen and Wegener, the Arctic Ocean

Fig. 40.10. Taylor's view of continental drift toward the equator. Reproduced from Van Waterschoot van der Gracht (1928).

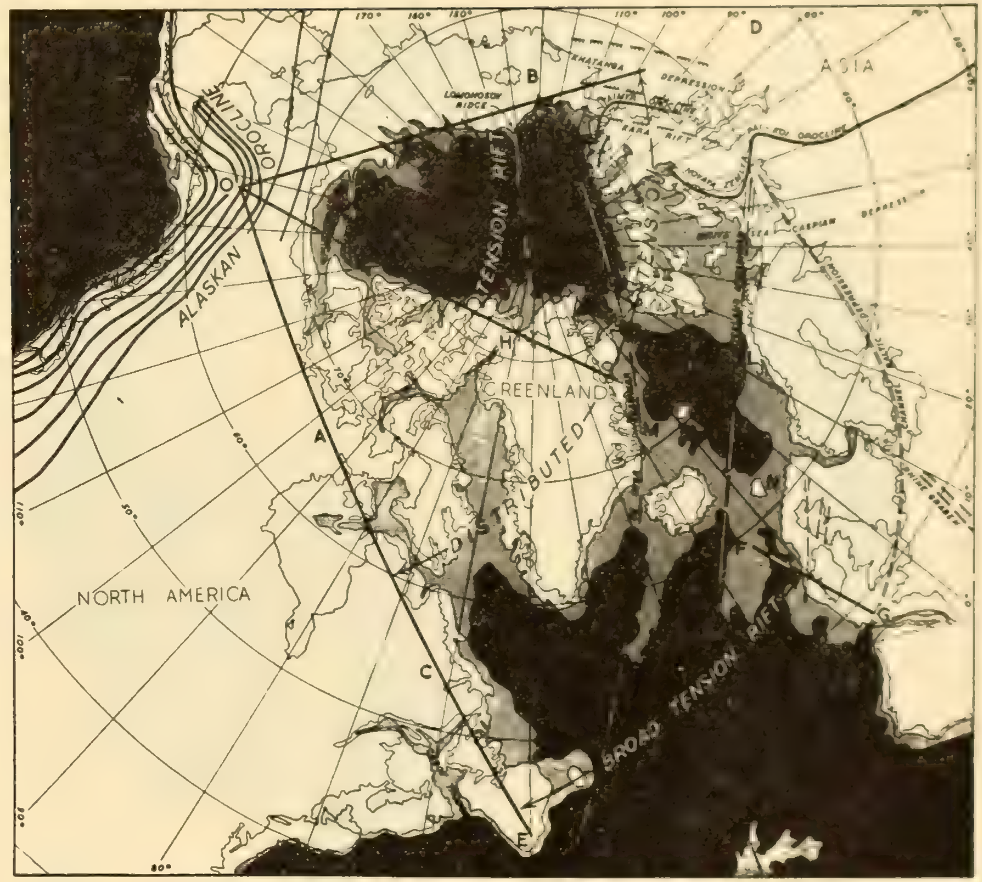


Fig. 40.11. Carey's (1958) concept of the origin of the Arctic basin.

basin is underlain by simatic crust. This is the seismic velocity layer considered today to be made up of a silicate of gabbroic composition.

In 1937 Du Toit presented a theory of origin of the Arctic and North Atlantic basins using the concept of continental drift facilitated by one or two major strike slip faults. The blocks bounding the Arctic basin are presumed to have rotated apart and the movement to have been accommodated by strike-slip along faults.

Elaborating on the ideas of Du Toit, Carey (1958, pp. 195–216) presents the following theory:

1. Scissors-like drifting apart to form the Arctic basin, hinging at a point in south central Alaska in what the present writer recognizes as the Nevadan orogenic belt. See Fig. 40.11.

2. The triangularly shaped basin is a tension rift with two sides being the radii from the hinge point and the third side a "megashear" or strike-

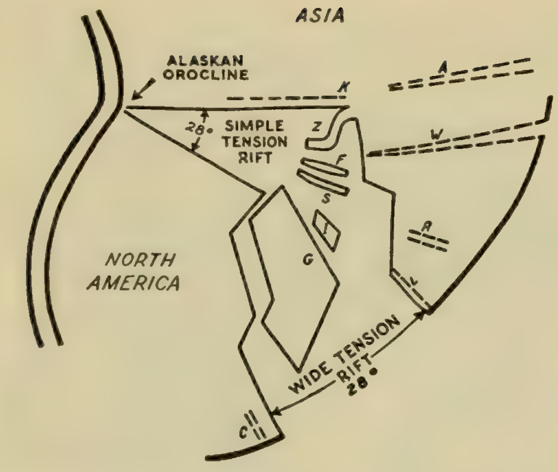


Fig. 40.12. Gross elments of orocline-sphenochasm concept of origin of Arctic basin. Simplified by omission of transcurrent movement. Reproduced from Carey (1958).

slip fault (Spitzbergen to Severnaya Zemlya). Such a tension rift he calls a sphenochasm.

3. The bending of the orogenic belts of the western American Cordillera in Alaska is believed to be the result of the rotation of the blocks which opened up the "Arctic sphenochasm." The bent segment of the orogenic belts is called the "Alaskan orocline."

4. The Lomonosov Range is believed to be a stretched-out more viscous part of the crust across the Arctic sphenochasm. The concept will be understood if the following model anology is considered. Quoting from Carey (1958, p. 195):

> If I break a slab of toffee which is cold and brittle except for one warm spot, the slab will break cleanly except at the hot spot where a thread of toffee will be drawn out across the rift. The thread will be straight or curved according to the path of separation. If the isotherm at which fracture passes into flow is locally above the Mohorovicic discontinuity sialic material will rise into the rhombochasm along with the rising mantel material and form a thread of sial across the rhombochasm. In view of the density difference it will endure permanently as a submarine ridge on the ocean floor. For such threads the name nematath [from Greek meaning, thread and stretched] is proposed. In practice I find that such nemataths commonly join similar igneous centres across the rhombochasm, giving support to the above hypothesis of their generation. Where the transition from fracture to flow is below the Mohorovicic discontinuity, no nematath results even if the isotherms are higher in some places than others.

5. The Novaya Zemlya, the Pai Khoi, and Severnaya Zemlya orogenic arc segments are considered oroclines and related to the "Iceland megashear." Restored to original positions they form a continuous, smoothly arcuate, Hercynian orogenic belt from the Urals to the North Greenland–Ellesmere orogenic belt. Norway would lie along side east Greenland and the Caledonian belts of each become parts of one broader, original zone. The Alaskan orocline is the hub of all movements of the northern hemisphere.

6. Paleomagnetic polar wandering is considered in light of the Alaskan orocline theory, and found to conform better to it than to other proposed patterns of fragmental shifts or drifting.

An idealized bold portrayal of the fracture and drift pattern of the Alaskan orocline and related features is shown in Fig. 40.12.

Rift Theory. Heezen, in November of 1956 (p. 1703), presented a paper at the meetings of the Geological Society of America in which the Mid–Atlantic rise and rift zone were postulated to extend to and across the Arctic basin. Since then papers by Heezen and Tharp (1959) and Ewing and Heezen (1957) have appeared which elaborate more on the concept. Figure 40.13 is a map supplied the writer by Dr. Heezen which shows the modern seismic activity of the Arctic and a new interpretation of the bottom topography of the Greenland Sea basin and the Arctic Ocean basin on the Eurasian side of the Lomonosov Range. He divides the deep basin (C) into two longitudinal parts with a gentle medial rise broken by a rift valley along the active seismic zone. The rift topography is born out by the sonic depth profile of the *Nautilus* (Fig. 40.7). The deep trench across Nansen's sill is along the postulated rift zone, and supports the concept of rifting, but it also favors drifting.

The gentle rise and medial rift constitute a tectonic element compatible with oceanic crust, and if similar to the Mid–Atlantic rise, we must postulate the zone to be one of volcanic activity.

Heezen postulates an expanding earth and the widening of the ocean basins as a result. The broad Mid–Atlantic rise and medial rift have developed progressively as the expansion occurred.

Conflicts and Problems. Assuming that the Arctic basin is underlain by oceanic crust, which seems probable, then the postulated belts across

it pose a problem. As far as the writer is aware no fold belts have thus far been proved in an ocean-type crust although parallel ridges and valleys have been taken to mean folding in one or two places. Generally, when fairly well defined, the parallel features are asymmetrical ridges or escarpments, and considered of fault origin. The continuity of fold belts across the Arctic basin, therefore, is to be considered doubtful.

The premise of fold belts across the basin was one of the chief reasons for postulating subsidence of part of a continental crust, but if the folds are doubtful and the crust is seismically oceanic, then the subsidence theory is improbable. A conflicting situation exists in Hakkel's publication (Hope, 1959a) in that he shows in a cross section oceanic crust, without any indication of or provision for folding, yet on the map he indicates a continuous fold belt across the deepest basin.

The theories of permanent ocean basins and of continental drift both provide for oceanic crust under the Arctic basin but in both the Lomonosov Range poses a problem. It does not appear to be of volcanic origin from the shape given it so far by contourers. If volcanic, the supporters of most any theory for the origin of the Arctic basin would find a compatible place for it in the framework of their concepts, but the nonvolcanic nature is a real enigma. Even the stretched-out nonorogenic thread idea of Carey is difficult to visualize without magmatism. If the future soundings indicate that the range is volcanic, and this is possible, then we will have been trying to solve nonexisting problems.

The rift theory, in view of the seismicity, seems attractive. It is in harmony with oceanic crust, but as far as the writer understands it does not present an explanation of the origin of the nonvolcanic (?) Lomonosov Range. The basins on the Alaskan side of the Range are not accounted for in the rift theory.

Both the subsidence theory and drift theory provide for a sourceland of sediments north of Alaska; the theory of permanence fails in this respect.

The orocline concept is complex but provocative and undoubtedly will elicit a good deal of attention in the future. More and better data are needed before further progress can be made on the origin of the Arctic basins. The published record to the time of this writing leaves the subject an enigma.

Fig. 40.13. Heezen's rift theory of the Nansen basin. Map kindly supplied by Dr. Heezen. The dots are earthquake epicenters.

41.

GULF COASTAL PLAIN

GENERAL CHARACTERISTICS*

Topography

The Gulf Coastal Plain is coextensive with the Atlantic Coastal Plain (discussed in Chapter 10) and, together, from Tampico, Mexico, to Cape Cod, Massachusetts, they are 3000 miles long. The Gulf Coastal Plain averages 250 miles wide, and the Mississippi embayment from the delta to Cairo is 575 miles long. The peninsula of Florida is 400 miles long. See Fig. 41.1. This vast plain rises very gently from the sea, and in parts of Texas attains an elevation somewhat more than 1000 feet. Beyond the Rio Grande in Mexico, the country that can properly be classed as coastal plain narrows toward the south and becomes structurally more complex than in the United States. At Tampico, it is very narrow and continues so to Yucatan, where the plain broadens to include most of the peninsula.

Geologically, the coastal plain extends out under the sea to the outer margin of the continental shelf.

* For an up-to-date detailed account, see Grover Murray's *Gulf Coastal Plain* (Harper & Brothers, New York, 1961).

Sedimentary Rocks

The Gulf Coastal Plain is underlain by a series of sedimentary formations composed chiefly of sand, clay, marl, limestone, and chalk, with subordinate amounts of salt, diatomaceous earth, volcanic tuff, and gravel. The calcareous deposits are more abundant in lower formations and along the seaward margin. The various sediments range in age from Late Jurassic to Recent and are mainly unconsolidated, though some indurated layers are intercalated from place to place. All the beds are lenticular and interfingered with others, and no two columnar sections are similar unless close together. This diversity in succession poses a constant problem for the stratigrapher, and microfossils have proved invaluable in correlation.

The Gulf Coastal Plain sediments were deposited in seas that invaded the margin of the continent. Several rivers draining the central part of the continent deposited vast amounts of sand, silt, and clay in the sea as the crust along the invaded margin subsided, and large amounts of chemical precipitates from the sea water were added. As a result, a great wedge was built up that thickens seaward. Along the site of the present coast of Mississippi, Louisiana, and Texas, the wedge of sediments is estimated to range from 20,000 to 30,000 feet thick. In spite of the great thickness, the wedge is very thin in relation to its length in cross section; and if it is laid out to true scale, one is impressed with the very small angle of tilt imparted to the beds by the subsiding of the land.

The stratigraphy of the central part of the Gulf Coastal Plain (Texas, Arkansas, Louisiana, Mississippi, and Alabama) is summarized in Figs. 41.2 and 41.3.

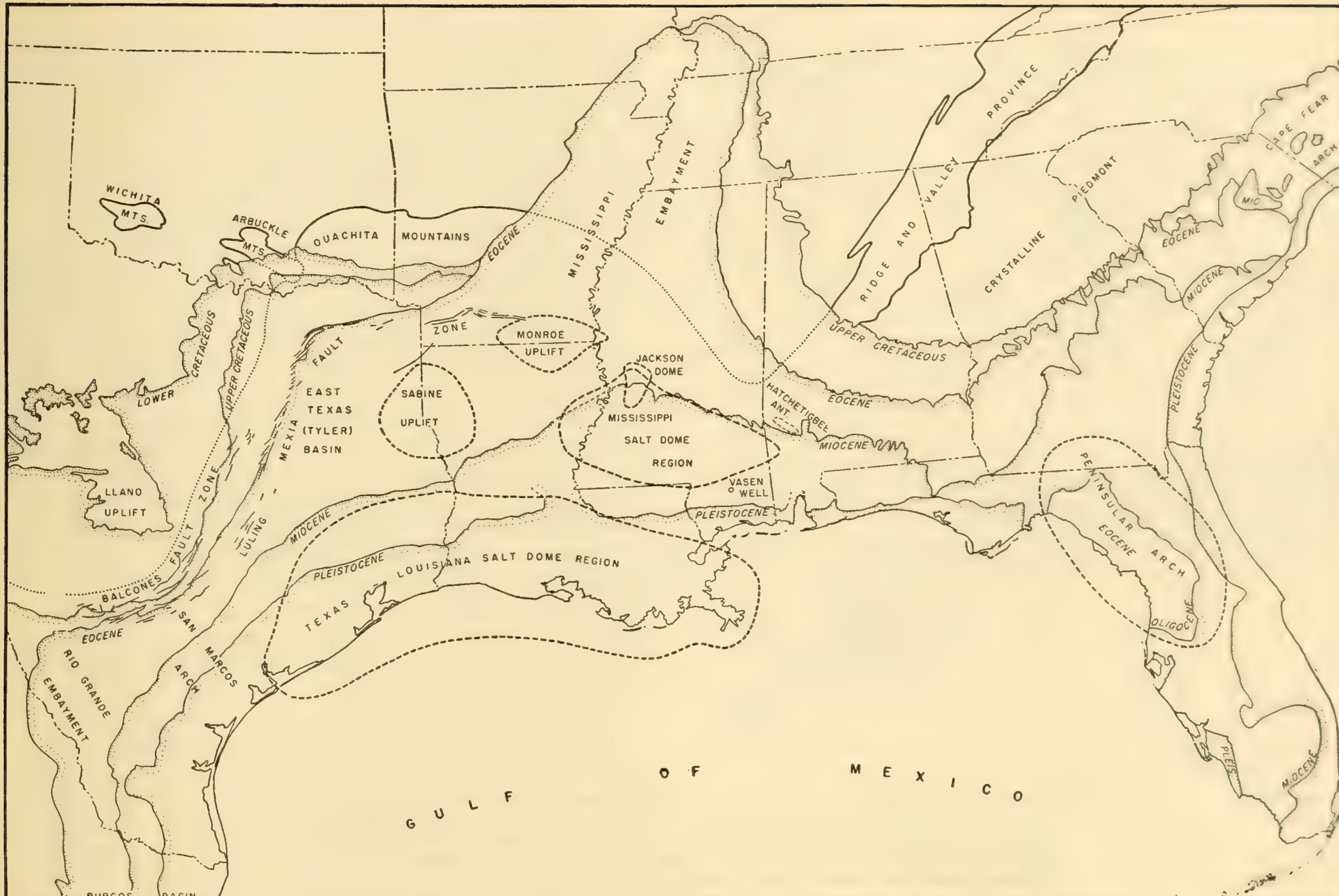


Fig. 41.1. Simplified geologic map of Gulf Coast. Structures enclosed by dashed lines are mostly subsurface; the Peninsular arch of Florida is a notable exception. Some salt domes have surface expression.

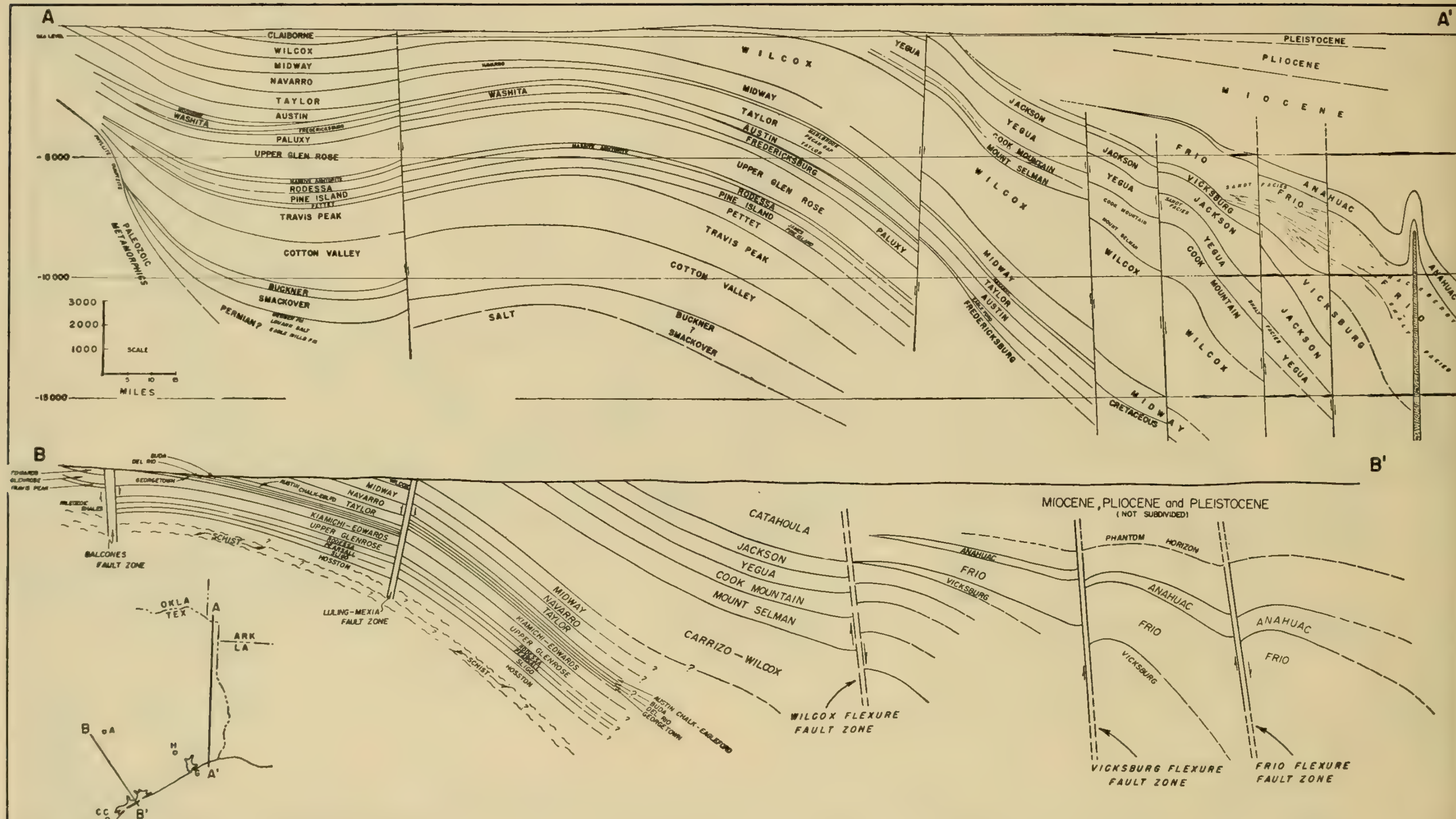


Fig. 41.2. Cross sections of the Gulf Coastal Plain in Texas. Taken from *Guidebook* for the Joint Annual Meeting of the A.A.P.G., S.E.P.M., and S.E.G. in Houston, 1953. Section A–A′ prepared by J. D. M. Williamson. Section B–B′ prepared by S. L. Stoneham.

Triassic sediments have not been recognized under the Coastal Plain, but during the Jurassic the Gulf waters invaded the continent and a succession of formations was deposited. From top to bottom they are the Cotton Valley, Buckner, and Smackover. Underlying the Smackover are the Werner gypsum and Louann salt formations which according to some authors are Permian (?), as in Fig. 41.2, and according to others Jurassic (Eagle Mills) as in Fig. 41.3. The Jurassic sediments were everywhere overlapped by the Cretaceous. The Lower Cretaceous sea and deposits extended across Texas and Oklahoma to connect with the vast epeiric sea of the Great Plains and the Rocky Mountains (see Plate 11). Upper Cretaceous seas probably spread over most of the Lower Cretaceous deposits in Texas, but their sediments have subsequently been stripped back so that the Lower Cretaceous now occurs farther inland (see Fig. 41.1).

The Upper Cretaceous deposits overlap the Lower in Mississippi, Alabama, Georgia, and South Carolina (see Fig. 41.3). After Late Cretaceous time the seas began a persistent retreat and the younger sediments are spread generally in successive belts toward the present Gulf of Mexico. Exceptions may be noted in the Mississippi embayment where on the west side the Eocene sediments overlap the Upper Cretaceous, and in Georgia and South Carolina where the Eocene sediments reach just beyond the Cretaceous in places and rest on the crystalline Piedmont.

The Cretaceous and Eocene seas especially extended up the Mississippi Valley, and their sediments reflect a transverse downwarp known as the Mississippi embayment. The evolution of the embayment is shown in four stages in Fig. 41.4.

The Rio Grande embayment is a gentle transverse downwarp and extends approximately from Corpus Christi northwestward for 200 miles up the Rio Grande. The axis of the downwarp lies somewhat northeast of the present river and close to the Nueces River.

The embayment is due strictly to Eocene downwarp, as only the Eocene sediments produce the embayed pattern. The Cretaceous strata cover large areas inland and merge with the widespread deposits of the Cretaceous seas in Mexico and the Rocky Mountains and Great Plains of the United States. See paleotectonic maps of Plates 11 and 12. The Oligocene, Miocene, and Pliocene deposits of the outer margin of the coastal plain continue around the Gulf without an embayment at the Rio Grande. A large part of the sediments of the Rio Grande area from Eocene to Pleistocene is of deltaic origin and was carried from the interior of the continent by rivers ancestral to the present Rio Grande, Pecos, and Nueces (Storm, 1945).

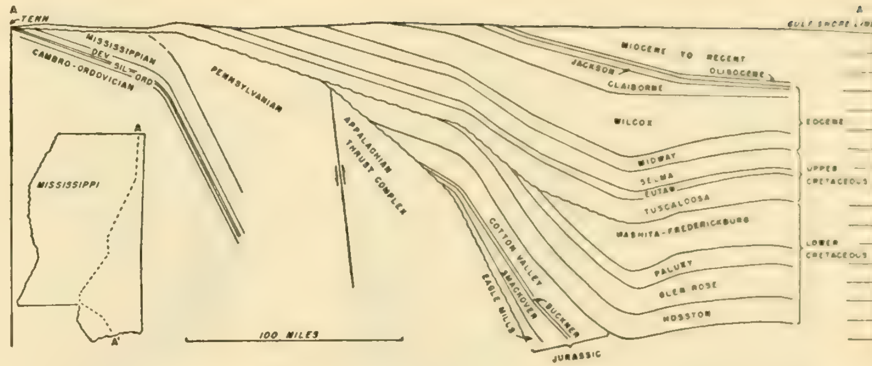
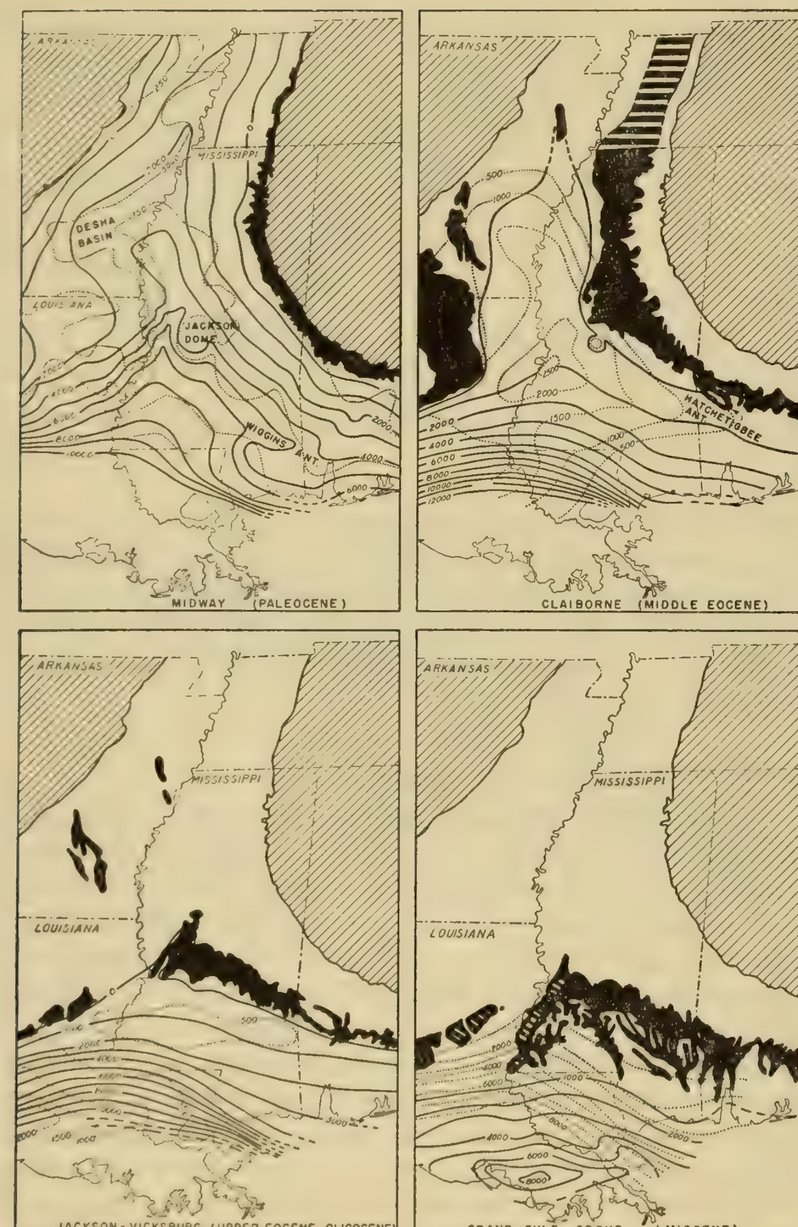


Fig. 41.3. Cross section of Gulf Coastal Plain through Mississippi, after Paul Weaver, 1951. Vertical scale is in thousands of feet.

Concept of the Gulf Coast Geosyncline

Recognizing the existence of 20,000 to 30,000 feet of sediments in the thick part of the wedge from surface and well studies of the coastal plain formations, and confirming the figures by geophysical data, Barton (1936) realized that the floor of the wedge was at least 10,000 feet below the floor of the Gulf of Mexico. In addition, he believed that the layer of sediments at the bottom of the Gulf of Mexico is only a few thousand feet thick at the most; and so he depicted a great elongate downwarp which he thought should rightfully be called a geosyncline. A number of papers by Barton and others have established the name Gulf Coast geosyncline firmly in the literature. The great accumulation of sediments along the sinking continental margin, however, has not yet been deformed—it has not been cast into folds and thrust sheets—but on the other hand, it is



one of very gentle seaward dips, except where locally disturbed by salt plugs, high-angle faults, and gentle warpings. A southward-migrating trough line has been postulated such that the maximum thicknesses of the different stratigraphic units are not superposed.

STRUCTURAL GEOLOGY

Balcones and Luling–Mexia Fault Zones

A complex assembly of faults follows approximately the border of the Tertiary and Cretaceous formations of the Coastal Plain in Texas. See map, Fig. 41.1. The zone of faults is located between the Edwards plateau of comparatively flat Cretaceous strata, and the seaward dipping Tertiary beds of the Coastal Plain. It crosses the San Marcos arch, which is a broad southeastward-plunging nose of the Llano dome.

The zone of faults has been divided in several ways. The *Tectonic Map of the United States* shows the southwestern part to be called the Balcones fault zone, and the northeastern part the Luling-Mexia. Weeks (1945) believes the Luling and Mexia are distinct and describes the three belts as follows:

Balcones Fault Zone.

Extending through the vicinities of the towns of Georgetown, Austin, San Marcos, New Braunfels, and some distance north of the towns of San Antonio, Hondo, and Uvalde is the Balcones fault zone with downthrown side principally on the southeast.

This zone of faults is located between the comparatively flat dip of the Edwards Plateau and the more steeply dipping beds of the Coastal Plain, and crosses the San Marcos arch, a broad nose which plunges southeastward from the Llano uplift. In the vicinity of Austin, Travis County, the total throw across the Balcones zone of faults approximates 900 feet; in northwestern Bexar County, 1200 feet; in northeastern Uvalde County, 500 feet; and in southwestern Uvalde County, 200 feet. In Kinney County the Balcones zone of faults dies out.

Fig. 41.4 Distribution, thickness, and structure contour maps of the Mississippi embayment and delta regions, after Murray, 1947. The black areas are the areas of outcrop, the solid lines are structure contours, and the dotted lines are isopachs.

Luling Fault Zone.

The Luling fault zone lies coastward from the Balcones zone and is composed principally of faults with downthrown side on the northwest. Examples of this zone are: (1) the Staples, Larremore (along which the Larremore oil field is located), and Lytton Springs line of faults in Guadalupe, Caldwell, and Bastrop counties; (2) the Luling fault in Guadalupe and Caldwell counties along which the Luling oil field is located, and which extends northeastward cross Caldwell County and into Bastrop County; (3) the Darst Creek–Salt Flat line of faults along which fields of these names are located in Guadalupe and Caldwell counties; and (4) the Somerset and Alta Vista faults in Atascosa and Bexar counties. All of these faults have considerable length. The average throw approximates 450 feet.

In Caldwell County along San Marcos River, a total throw of more than 1,500 feet is indicated on faults of the Luling zone. The faults of this zone have the greatest throw and are most numerous from Travis and Bastrop counties southwest through Bexar County, thus crossing the San Marcos arch.

Mexia Fault Zone.

Farther down the dip than the fault zones described above is the Mexia zone of faults characterized by faults with downthrown side on the southeast and by faults with downthrown side on the northwest. Both faults commonly occur together with a graben of varying width between them. In the Mexia area, the name Tehuacana has been given to faults on the northwest side of the graben. The Mexia zone of faults extends from the vicinity of Mexia, Limestone County, northeastward and eastward around Tyler basin. Faults in southwestern Arkansas probably are a part of this zone. From Mexia southwestward this zone of faults extends far into South Texas and offsets down the dip at various points. At Mexia, Midway beds at the surface are cut by the faults; in Lee County, Mount Selman; in Bastrop and Fayette counties, Cook Mountain and Yegua; and in Gonzales County, Yegua and Jackson.

The zone of faults extending southwest through parts of DeWitt, Karnes, Goliad, Bee, Live Oak, and Duval counties may be part of this zone. Representative faults are those along which oil and gas have accumulated in northern Bee County in the vicinity of Pettus. Considering all of these faults as belonging to the Mexia zone, in Texas alone the length of the zone is over 500 miles.

In the region of Mexia many of the faults along which oil and gas production is obtained from the Woodbine are *en echelon*, with the south end of the fault at the north passing west of the north end of the fault at the south. This causes closure in this direction. There is a structural high in the region of Mexia, and south of this high the strike of the beds and the strike of the faults tend to converge at the south end of each fault structure and tend to divirge at the north end. This lack of effective north closure, plus absence of Woodbine sand, probably is the reason for barren structures on the south toward the Falls County regional low.

Minor movements may have occurred in Cretaceous time, but the main displacements came in late Oligocene (late Catahoula) or Miocene (early Oakville) time (Weeks, 1945). The sediments of the Catahoula and Oakville reflect the movements. Because certain Pliocene beds are displaced less than older beds, it follows that some movement in places has occurred in post-Pliocene time.

The structure of the Coastal Plain from the Bend arch of central Texas eastward across the fault zones to the Sabine uplift is shown in the cross section of Fig. 41.5.

Flexure Fault Zones

Paralleling the coast of Texas and shoreward of the Miocene boundary (Fig. 41.1) are three flexure and fault zones. These are called flexure zones or flexure fault zones, and they are shown in the lower cross section of Fig. 41.2. The Gulf side is down 500 to 1500 feet, but the unusual aspect is the reverse (?) drag aspect of the beds on the downthrown side. This has been interpreted as sagging or slumping of the beds incident to the tendency of fissure opening as down-tilting of the block toward the Gulf occurs. The faults die out upward in the Miocene and Pliocene sediments and hence are about mid-Tertiary in age. Needless to say the flexure fault zones are the sites of very productive belts of oil fields.

Tyler or East Texas Basin

East of the Balcones and Mexia fault zones is the Tyler basin, so called on the *Tectonic Map of the United States*, but often named the East Texas basin. See cross section of Fig. 41.5. It is the result of gentle dips eastward off the Bend arch of central Texas and westward off the Sabine uplift. It consists of a thick Tertiary and Upper Cretaceous sequence of beds. The Lower Cretaceous succession has not been penetrated in the deeper parts of the basin, nor has the mother salt that has spawned a score of salt domes within the basin.

Sabine and Monroe Uplifts

A large gentle dome in easternmost Texas and northwestern Louisiana is reflected in the surficial Tertiary strata, and is known as the Sabine

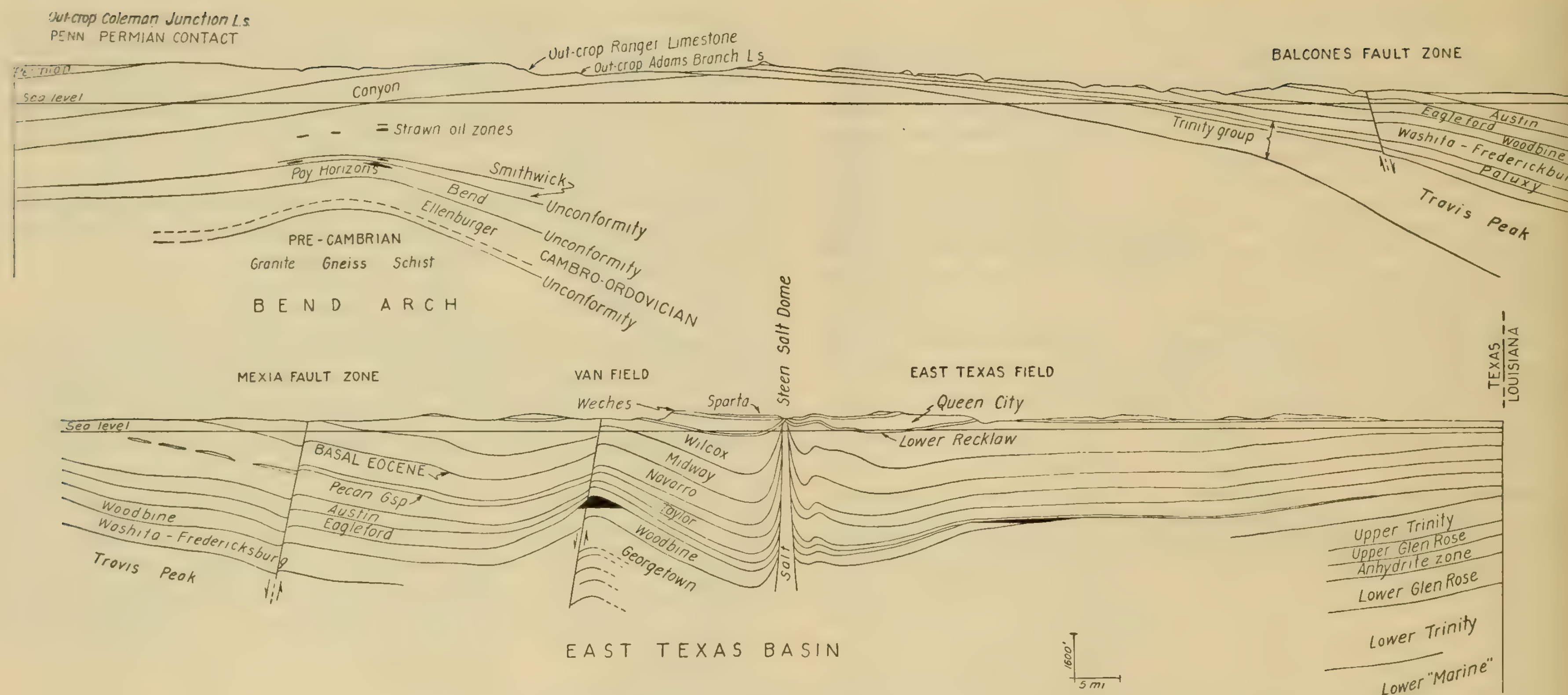


Fig. 41.5. Cross section from the Bend arch, central Texas, eastward to the Sabine uplift.

uplift. The shallow structural sag on the west is the Tyler basin, just described. A small and shallow syncline separates the Sabine uplift from the gentle Monroe uplift in northern Louisiana and southern Arkansas. The axis of each uplift trends northwest-southeast. The doming started in Cretaceous time. The Sabine uplift was an island at the close of the Early Cretaceous, and the Monroe uplift was an island during much of Late Cretaceous time. The Sabine uplift especially was effected by upward movements in post-Claiborne (post-middle Eocene) time, and this doming with ensuing erosion has left a core of Midway (Paleocene), Wilcox (early Eocene), and Claiborne (middle Eocene) sediments surrounded by younger formations (Murray and Thomas, 1945). See Fig. 41.6.

Jackson Dome

The Jackson dome is a sharp uplift in the subsurface in west-central Louisiana. See Fig. 41.7. It is about 30 miles across. Local doming sufficient upon erosion to expose the Upper Jurassic Cotton Valley formation occurred at the close of the Late Cretaceous. The amplitude of the fold is about 10,000 feet, but the dips shown on Fig. 41.7 are excessive owing to the grossly exaggerated vertical scale.

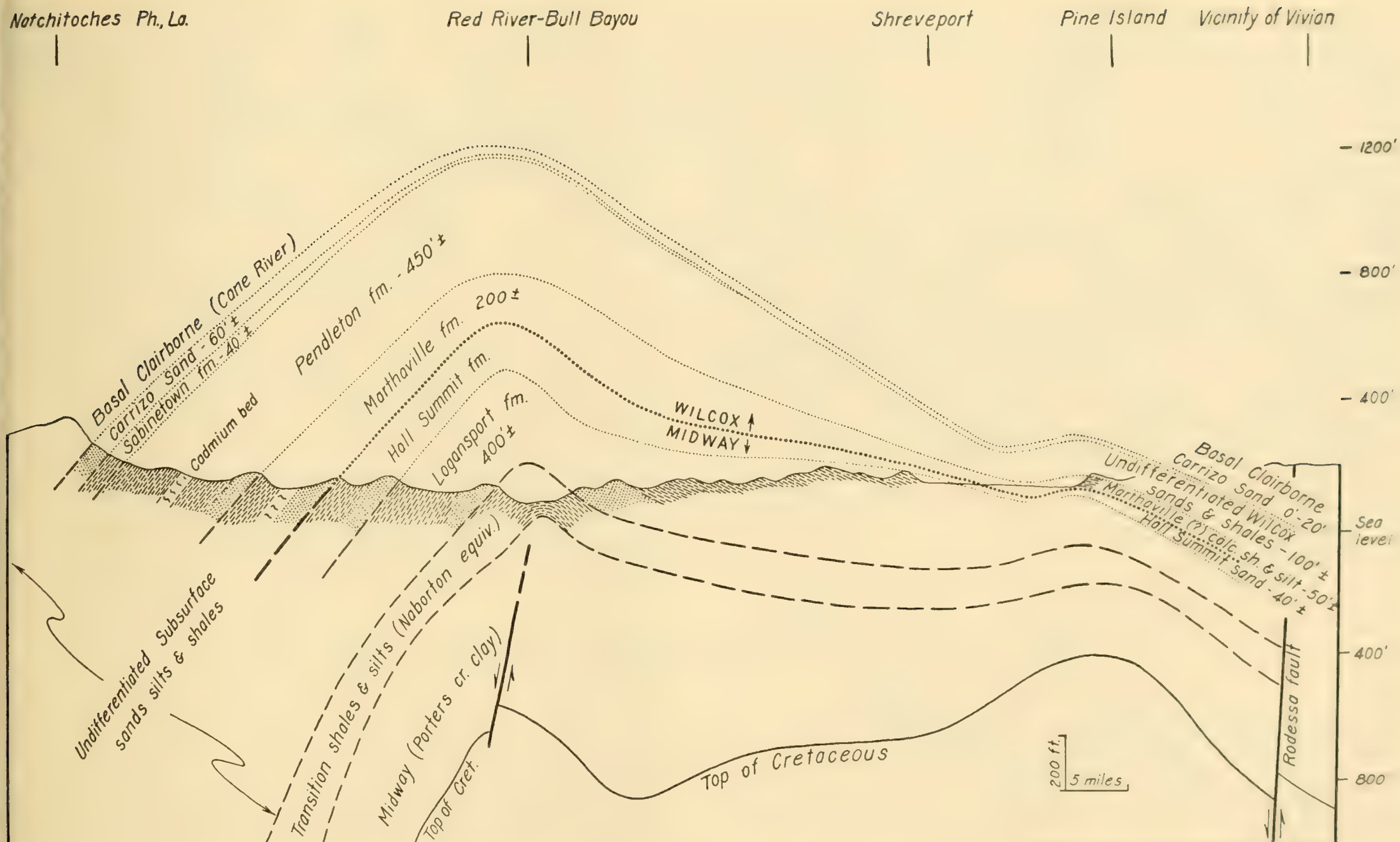


Fig. 41.6. Cross section of the Sabine uplift approximately from north to south showing the details of the Tertiary stratigraphy. The vertical scale and hence the structure are tremendously exaggerated. After Murray and Thomas, 1945. Midway is Paleocene in age, Wilcox is lower Eocene, and Claiborne is middle Eocene.

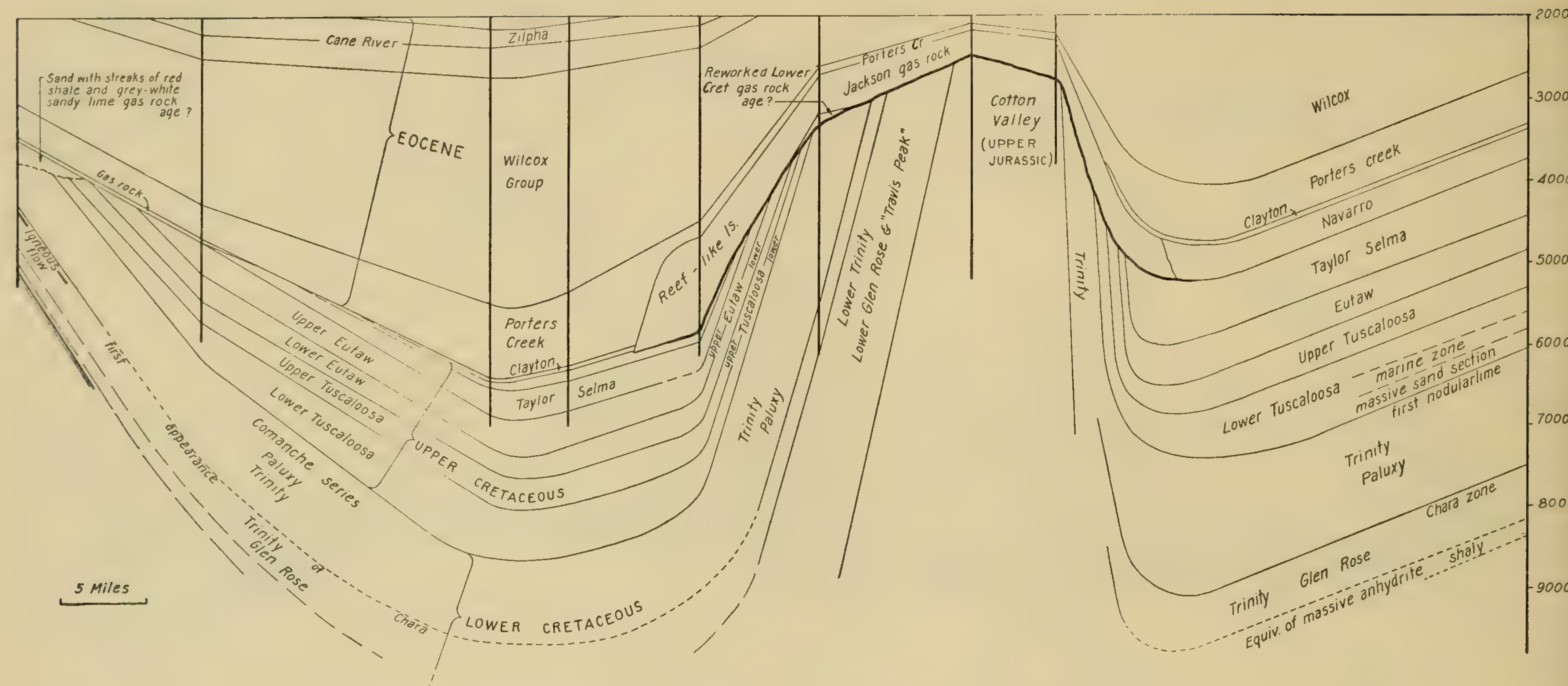


Fig. 41.7. Cross section through the Jackson dome, Mississippi, taken from west to east. After McGlothlin, 1944.

Another structure in the Mississippi embayment is the Desha basin north of the Monroe uplift. It is not a closed basin but opens on the east into the broad embayment. In southwestern Alabama is the Hatchetigbee anticline which trends to the northwest and has surface outcrop expression.

Salt Domes

Distribution. Semicylindrical masses of salt have thrust their way upward in the poorly consolidated sediments of the Gulf Coastal Plain in a variety of forms. They are known as salt domes or salt plugs, characteristically from one-half to two miles in diameter, and are the loci of many fine oil fields. Over 200 are now known in the Gulf Coast. They are distributed in two general groups: (1) the coastal domes principally through southern Texas, the Mississippi delta of Louisiana, and the shallow offshore shelf (Figs. 41.1 and 41.8); and (2) the interior domes. Some coastal domes also occur in northern Mexico in the Vera Cruz-Tabasco basin. The black dots on Fig. 41.9 indicate the salt domes discovered to date. It will be seen that the greatest number are in the

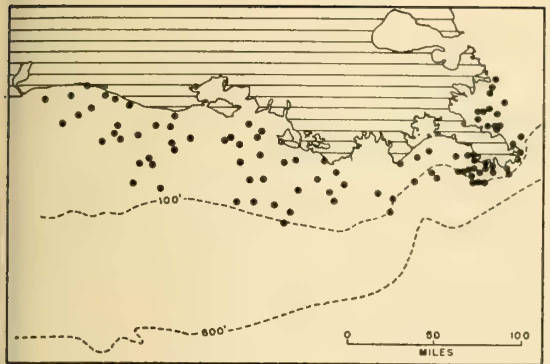


Fig. 41.8. Offshore salt domes on the continental shelf of Louisiana as of March, 1958. After Habarta, 1958.

coast belt. Those in the interior are divided into three areas, one in the Tyler basin, one in the eastern part of the Sabine uplift, and one in a broad zone across south central Mississippi.

Classification. Salt domes are classified in several ways. The divisions deep, intermediate, and shallow are the most commonly mentioned. Deep domes are considered those whose salt core tops are greater than 5000 or 6000 feet below the surface (Billings, 1942), intermediate domes between 6000 or 5000 and 3000 or 2000, and shallow domes, less than 3000 or 2000 feet deep. Some have reached the surface. Deep domes are divided into those whose salt has been reached by the drill and those whose salt is below any deep wells.

Another classification concerns the relation of the salt plug to the country rock. If the salt has simply domed the overlying beds in the manner of a concordant laccolith, the structure is called a nonpiercement dome. If, on the other hand, the salt has penetrated through the beds, the structure is said to be a piercement dome. Generally all domes are now considered as piercement type, whether shallow, intermediate, or deep-seated. Refer to Fig. 41.9 illustrating the origin of salt domes for these types and also a number of transitional ones.

Some salt domes have mushroomed out at the top, and the cap rock and part of the salt core is said to overhang. These horizontal expansions or wedges have been drilled through and their presence thus demonstrated.

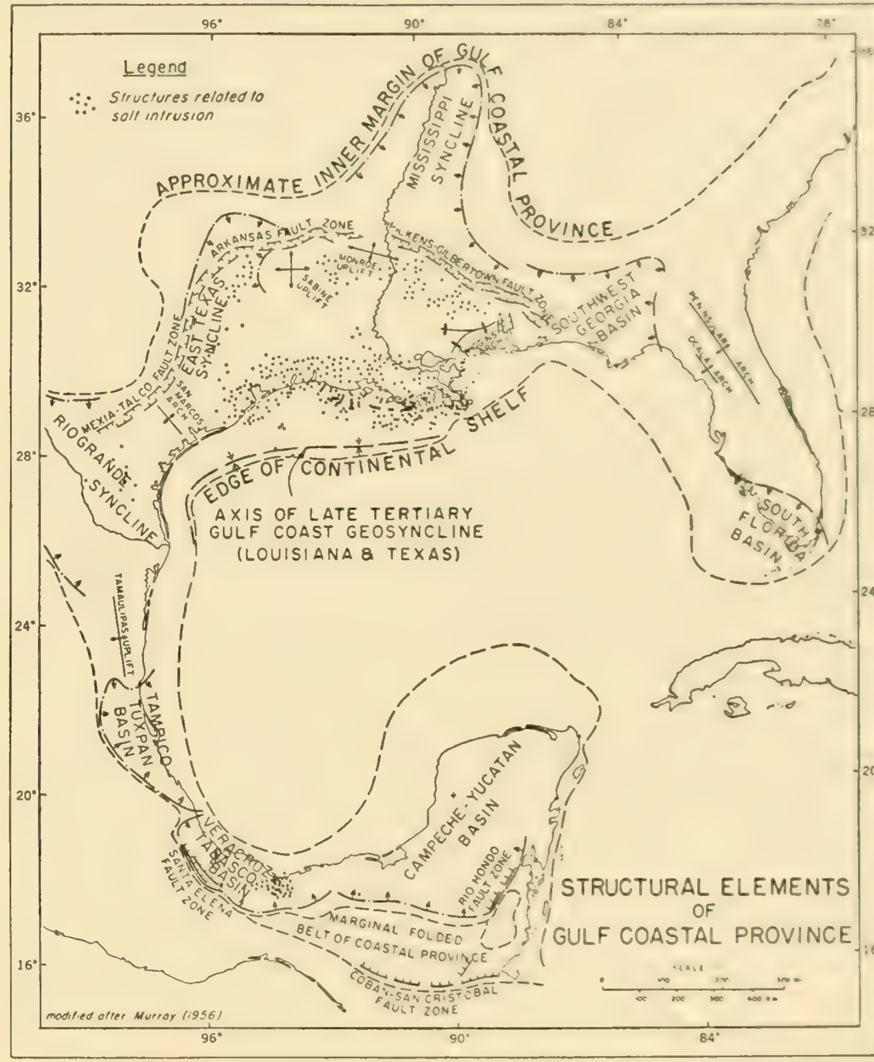


Fig. 41.9. Structures of the Coastal Plain around the Gulf of Mexico. Reproduced from Atwater and Forman (1959).

Salt domes also differ on the basis of their cap rock. Overlying the massive rock salt of the core of the dome is an irregular layer or cap of limestone, gypsum, and anhydrite. Limestone is generally at the top, and anhydrite at the bottom of the capping layer. The cap rock of a few domes contains immensely valuable native sulfur deposits.

The sedimentary layers over the salt domes have been domed up gently, and the layers adjacent to the intrusive salt plugs have been dragged upward to a greater or lesser extent. Oil is found, therefore, over the salt plug in the domed strata, on the flanks against the plug, under overhangs, and in associated fault traps.

Deep drilling of certain salt domes of southern Louisiana shows that very large volumes of salt are involved, and that one structure at a depth of 20,000 feet has an area of 200 square miles, and contains 265 cubic miles of salt above the 20,000 foot datum (Atwater and Forman, 1959). Further, the salt is intrusive like major igneous discordant plutons; the country rock has been "replaced" rather than shoved aside. The manner of emplacement is an enigma. Also, large masses of contorted shale have been carried up far above their normal stratigraphic position and look like intrusive masses themselves. The intrusive action has been localized to one part of the large dome at one time, and then to another part at another time.

Origin. Salt domes result from the plastic intrusion of sedimentary rock salt into overlying beds. Rock salt under pressure deforms easily and flows from places of greater pressure to places of lesser pressure. Many geologic observations confirm the concept that salt flows easily and that the associated shales are commonly intensely deformed. According to Nettleton (1936) it is reasonable to assume that the present shape of the dome is due to (1) initial configuration that localized the dome, (2) the thickness of the mother salt layer, (3) the strength or viscosity of the overlying rocks, and (4) the strength or viscosity of the salt. Figure 41.10 shows the theoretical development of salt domes under three conditions.

A number of deep-seated salt domes are marked by faults that cut and offset the arched beds over the salt core. Perhaps the faulting is a general characteristic. The faults are normal and form a complex graben through the central part of the dome. Wallace (1944) believes that the common fault patterns are simple offsets and simple and complex graben such as illustrated in Fig. 41.11. The first fault that occurs is called the principal fault, which produces the simple offset. The next is the complementary fault (also called minor fault), and the next is another minor fault. The generalized diagrams of Fig. 41.11 give the impression that all graben cut across the domes; but as more is learned of the detail of the deep domes, more faults are recognized, and their ground pattern may be somewhat concentric in certain domes, somewhat radial in others, and crosscutting in still others.

The intruding salt has also caused small reverse faults on the sides of certain domes. These are significant in forming oil traps (Halbouty and Hardin, 1954, 1956).

Wiggins Anticline and the Deep Wells

Just 50 miles north of the Gulf of Mexico in southern Mississippi a well was drilled 20,450 feet deep in a subsurface structure called the Wiggins anticline. It is known as the George Vasen's Fee well and was completed in 1951 (Applin and Applin, 1953). At the total depth it reached rock salt of pre-Smackover (Jurassic) age. Nearly 5500 feet of consecutive cores of unmetamorphosed Jurassic strata were obtained. From a depth of 14,670 to the bottom the formations penetrated have been identified as follows:

Lower Cretaceous	
Lower part of Hosston fm.	275 feet
Dark brownish-red shale	
Upper Jurassic	
Cotton Valley group	2053 feet
Mostly nonmarine or detaic deposits in upper part; lower fourth is marine and fossiliferous	
Cotton Valley (?) group and Buckner (?) fm.	1700 feet
Nonmarine sandstone and shale	
Smackover formation	105 feet
Dark sandstone, siltstone, and shale	
Dips 25° to 60°	

Jurassic (undifferentiated)	
Smackover (?)	1620 feet
Limestone, dolomite, anhydrite	
Pre-Smackover	30 feet
Gray crystalline anhydrite and at bottom 1 foot of clear white rock salt	

A well near the front of the Mississippi delta penetrated to a depth of 22,570 feet and ended in Miocene strata (Paul Lyons, personal communication). In relation to the Vasen's well on the Wiggins anticline, 120 miles to the north, a marked southward dip is evident, which is reported as 7°. The steepening of southward dip in the Mississippi delta is prominent on the maps of Fig. 41.4, and for the Miocene beds a trough axis has been discerned extending east-west through the delta.

IGNEOUS ROCKS

Moody (1949) has summarized the igneous rocks of the Gulf Coastal Plain, both pre-Cretaceous and Cretaceous in age. The greatest concentration of igneous activity centers in the Monroe uplift and Jackson dome areas (as designated on the *Tectonic Map of the United States*) in the tristate area of Arkansas, Louisiana, and Mississippi. There, in well-drilling operations, alkaline and ultrabasic intrusive rocks have been drilled into, and also volcanic rocks in the form of flows or sills; pyroclastics are abundant, both alkaline and basic. Some of the intrusive bodies, possibly dikes and stocks, are definitely intrusive into Upper Jurassic strata. Some are older and believed to be related to the Triassic diabase of the Atlantic piedmont.

Throughout the entire northern part of the Coastal Plain in the Upper Cretaceous sediments fragments of volcanic rocks are found in association with the common sedimentary detrital minerals.

TAMPICO REGION, MEXICO

The Tampico region has a somewhat different Cretaceous geology from the rest of the Gulf Coast, but a similar Tertiary. Instead of an overlap from the Gulf, the Cretaceous beds are continuous with those of the interior Mexican geosyncline and the Parras basin. The Cretaceous beds of the geosyncline are intensely folded, and along the east front of the Sierra Madre Oriental they are thrust eastward in places. The zone from the Sierra Madre front to the coast, 60 to 100 miles wide, may be regarded as the coastal plain where the sedimentary rocks are fairly flat; but several anticlinal mountains (or hills) formed of Cretaceous rock interrupt the plain. The Tertiary sediments were deposited in seas that invaded the coast from the Gulf and buried unconformably a number of relief features.

The anticlinal or domal mountains that rise from the plain are, from north to south, the Sierra Burro, Lomerio Peyotes, Sierra Lampazos, Sierra San Carlos, and Sierra Tamaulipas (Muir, 1936). See maps, Fig. 42.1 and 35.1. The Sierra San Carlos has already been described in Chapter 28 and is fairly representative of the mountains east of the Sierra Madre front. Some of the ranges have gentle dips on the flanks from 3 to 10 degrees. The Sierra Papagayos is steeply folded, with dips up to 40 degrees and more. The doming of the Sierra San Carlos has been accentuated by the intrusion of a stock of nepheline syenite, and the folding of the Sierra Picochos has been influenced by intrusions (Muir, 1936).

All the ranges just mentioned in the coastal plain are parts of a continuous structural element and hence related genetically. The Sierra Tamaulipas anticline plunges southward, and the so-called northern oil fields are on its prolongation. See cross section of Fig. 41.12. Near the termination of the Sierra Tamaulipas on the southern flank is an offshoot named the Sierra de Buenavista. The Tamaulipas limestone in the core is intruded by a laccolith. Muir concludes that the forces that produced the mountains and oil-field structures of the coastal plain are due to vertically acting forces, in contrast to the Sierra Madre and interior structures which are due to horizontally acting forces.

The northern oil fields are in an area of Cretaceous rock that reaches nearly to the coast at Tampico. Immediately south of Tampico, beds of Eocene, Oligocene, and Miocene age lap 50 miles inland across the Cretaceous and bury an arcuate ridge which lies west of Tuxpam. Albian-Cenomanian reef limestones were probably laid down on a late Aptian

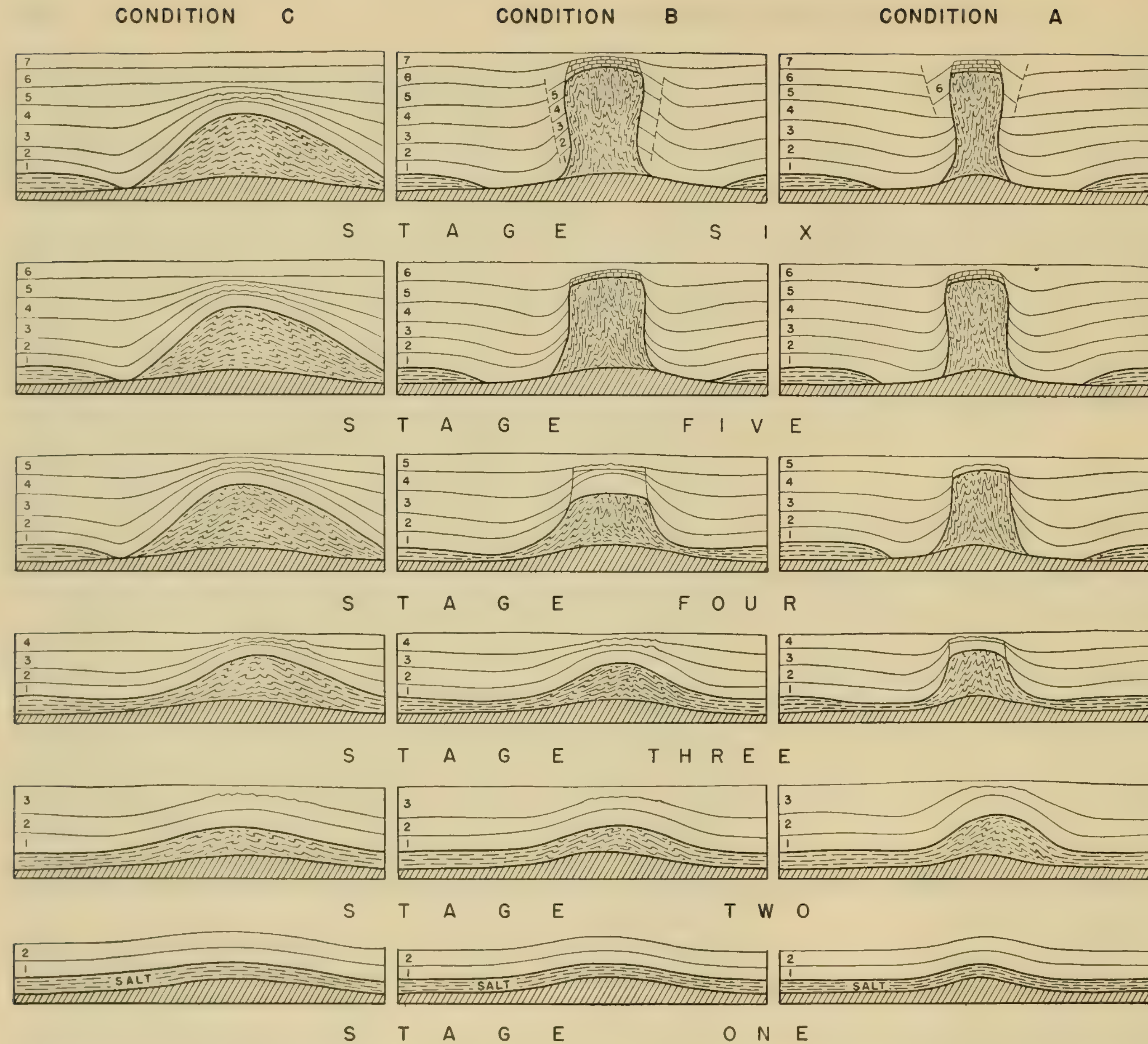


Fig. 41.10. Theoretical development of salt domes under various conditions, after Nettleton, 1936. Diagrams are patterned partly after model experiments involving viscous flow, and partly after actual examples.

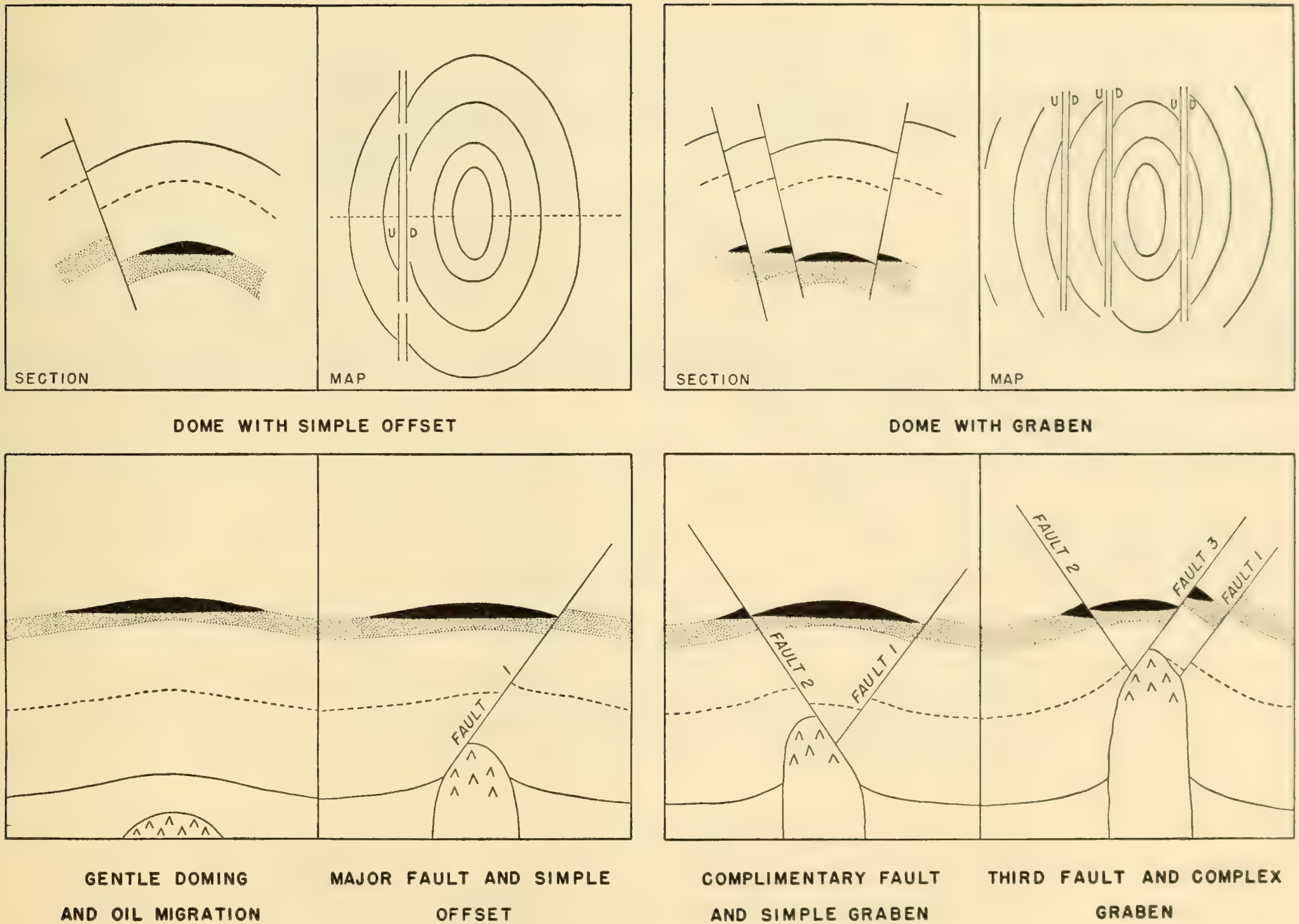


Fig. 41.11. Origin of faults over deep salt domes, after Wallace, 1944.

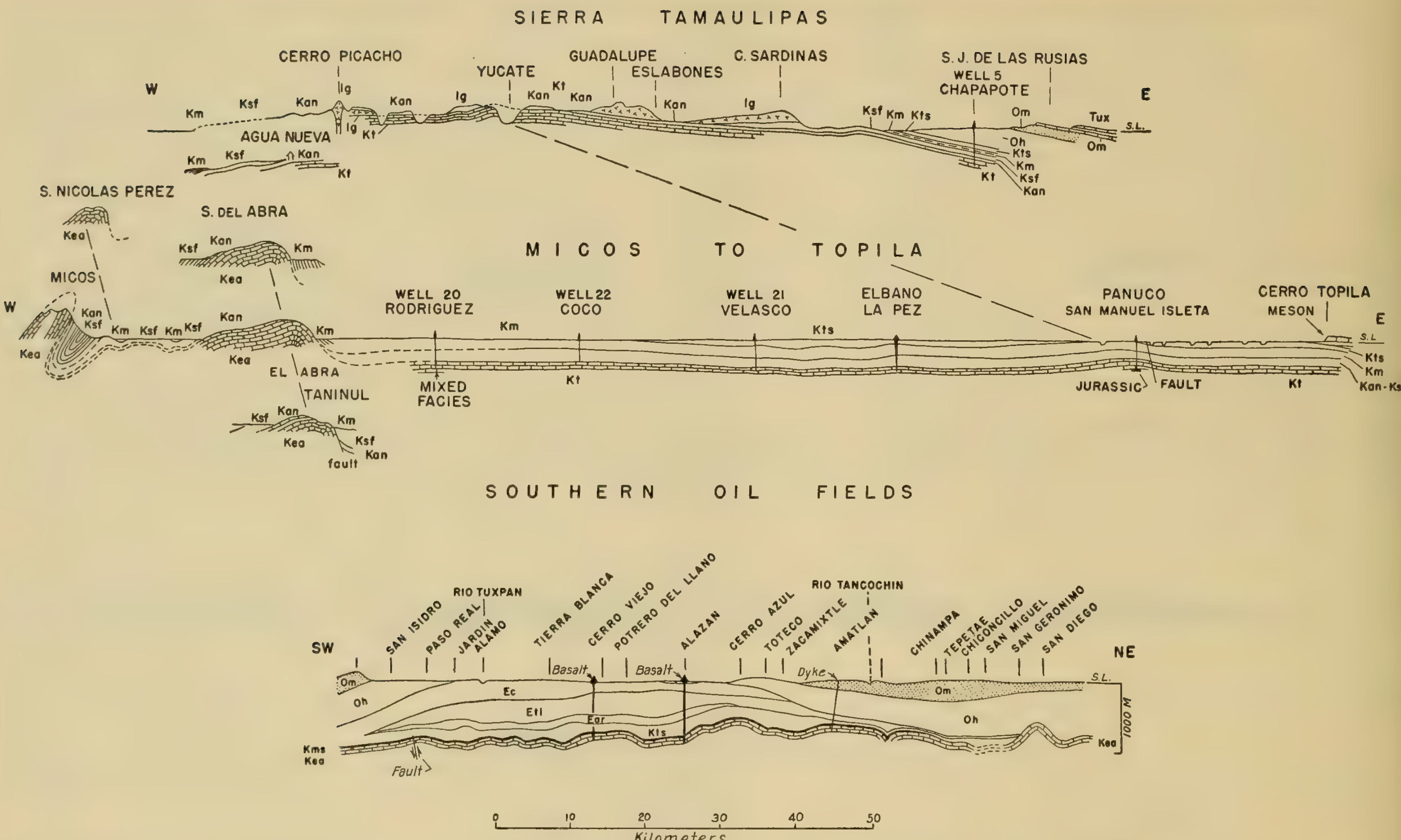


Fig. 41.12. Cross sections of the coastal plain of eastern Mexico after Muir, 1936. The upper section is the southward plunging end of the Sierra Tamaulipas, where the northern oil fields are located. The lower section is west of Tuxpam, and runs nearly north-south, longitudinal of the structure.

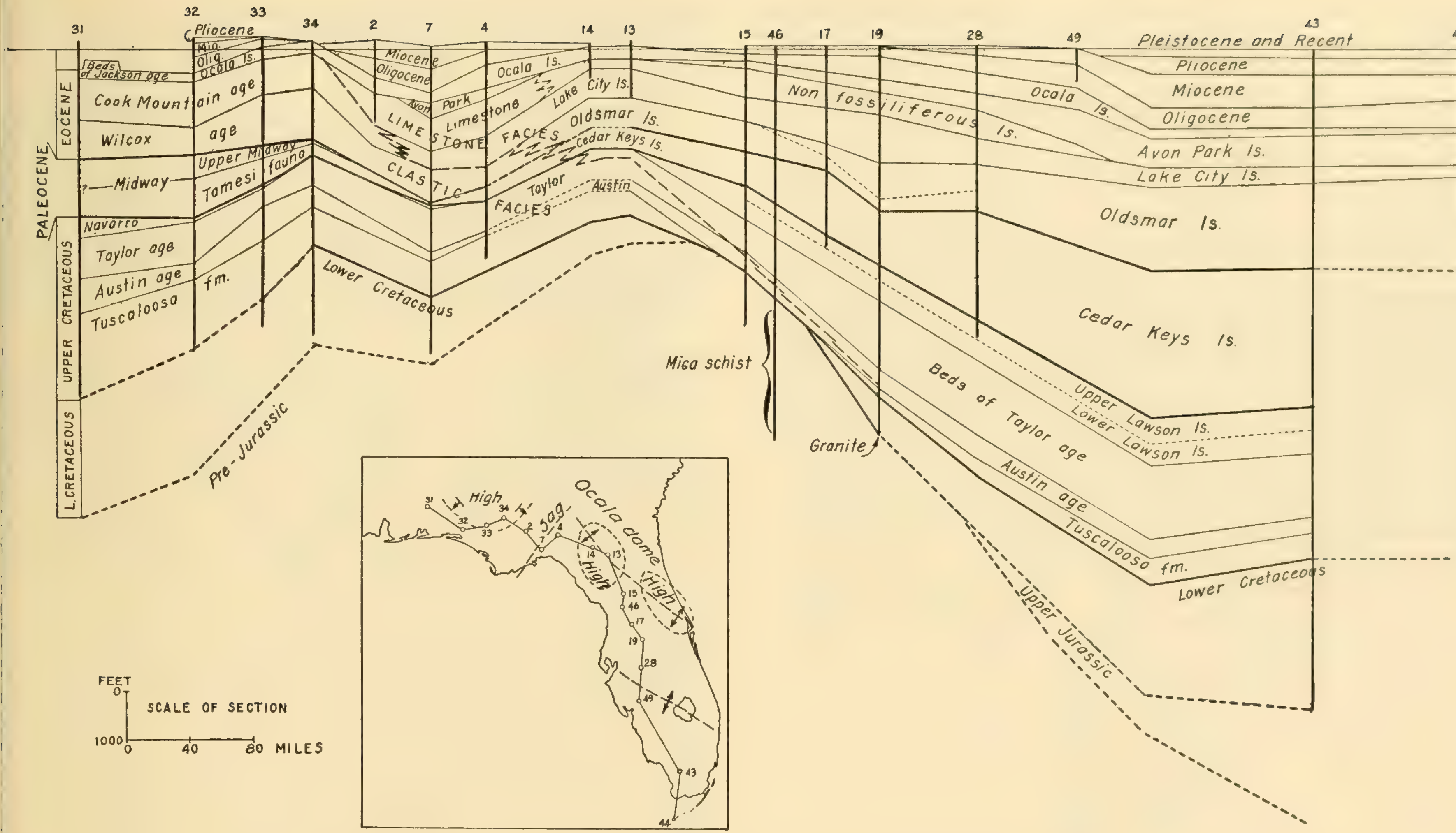


Fig. 41.13. Cross section from southern Georgia to Key West in southernmost Florida, after Applin and Applin, 1944.

submarine limestone reef that formed a ridge. The ridge area was elevated in post-Turonian time, and again repeatedly in Eocene and Oligocene time. With sedimentation repeatedly burying it, a number of unconformities occur around and over it. The spatial aspect of the buried ridge is difficult to fit into the structural picture (Muir, 1936).

FLORIDA PLATFORM

Sediments from Recent to Early Cretaceous age are known to overlie a crystalline basement in Florida, and beds are probably present in places between the Lower Cretaceous and the crystalline rocks. The thickness of the sedimentary cover ranges from 4350 feet in southeastern Georgia to more than 11,600 feet in the southern end of the peninsula. The earliest tangible history of Florida is that of the Early Cretaceous, when approximately the western half of the peninsula was submerged and the eastern half was land. A number of drill holes bear out this picture fairly well. If an outer belt of Lower Cretaceous strata in Georgia and the Atlantic Coastal Plain under the Upper Cretaceous and Tertiary beds is related to the land area of Florida, a long peninsula seems to have existed then, as now, only slightly eastward of the present.

If the isopachs of the Upper Jurassic of the Mississippi embayment region and Arkansas and Texas are projected to Florida where problematical Upper Jurassic has been recognized in just two deep wells (Applin and Applin, 1944), only the southern third of Florida seems to have been under water, and the rest was land. In fact a very broad land projection seems to have existed. See paleotectonic map of Plate 10.

The cross section and map of Fig. 41.13 show the stratigraphic and structural relations recognized in Florida. The chief structural feature is the Peninsular arch in the north-central part of the peninsula which first appeared in the Late Cretaceous. The axis of the arch trends northwestward and is parallel with a deep trough that centered in the Greater Antilles. The arch is also pronounced in the Middle and Upper Eocene beds, but with variations in detail. A flexure developed on the west flank of the Peninsular arch has distinct outcrop expression and is properly called the Ocala uplift, according to the Florida Geological Survey, but the large arch itself is commonly called the Ocala.

According to Applin and Applin (1944) the chief structural features of Florida are:

(1) An axis extending northwest from about Cape Canaveral on the east coast of Florida to south-central Georgia, upon which are located two large locally high areas; (2) a channel or trough extending southwestward across Georgia through the Tallahassee area of Florida to the Gulf of Mexico, nearly at right angles to the aforementioned axis; (3) an upwarped area in the vicinity of Jackson County, Florida, with dips extending away from it toward the southeast, south, and southwest; (4) a structurally low area with an axis extending northwest from the vicinity of Lake Okeechobee toward Tampa, approximately parallel with the axis first mentioned; (5) a possible second north-west-trending upwarped area at the south end of the Peninsula.

The modern peninsula of Florida is about the emergent third of a broad platform, as may be seen in Fig. 42.1 The shelf on the west side is 100 miles wide and ends in a very steep escarpment which carries down to the abyssal plain of the Gulf of Mexico. This West Florida escarpment has been thought of as a fault scarp (Jordan, 1951), but on hand of a uniform magnetic intensity field over the escarpment and the aseismic nature of the region, Miller and Ewing (1956) believe it is not due to faulting but to processes of sedimentation. The constitution of the crust under the Gulf of Mexico, and the origin of the Gulf will be discussed under a later heading.

The shelf on the east of the peninsula of Florida is continuous with and supports the Bahama Banks whose geology will be discussed in the next chapter. The great shelf region is largely one of carbonate deposition today, and as explained, has so been in the southern half since at least Early Cretaceous time. Accumulation has equaled subsidence, and the imposing submarine escarpments may be due to the growth of reefs, firm enough to keep the sediments from slumping down to the abyssal plain. Some local magnetic anomalies on the West Florida escarpment may indicate buried volcanic piles (Miller and Ewing, 1956).

A number of wells have penetrated the Mesozoic sedimentary rocks, and maps of the surface are shown in Fig. 41.14. The contour of the surface is that of the dominant Peninsular arch. The outcrop pattern, how-

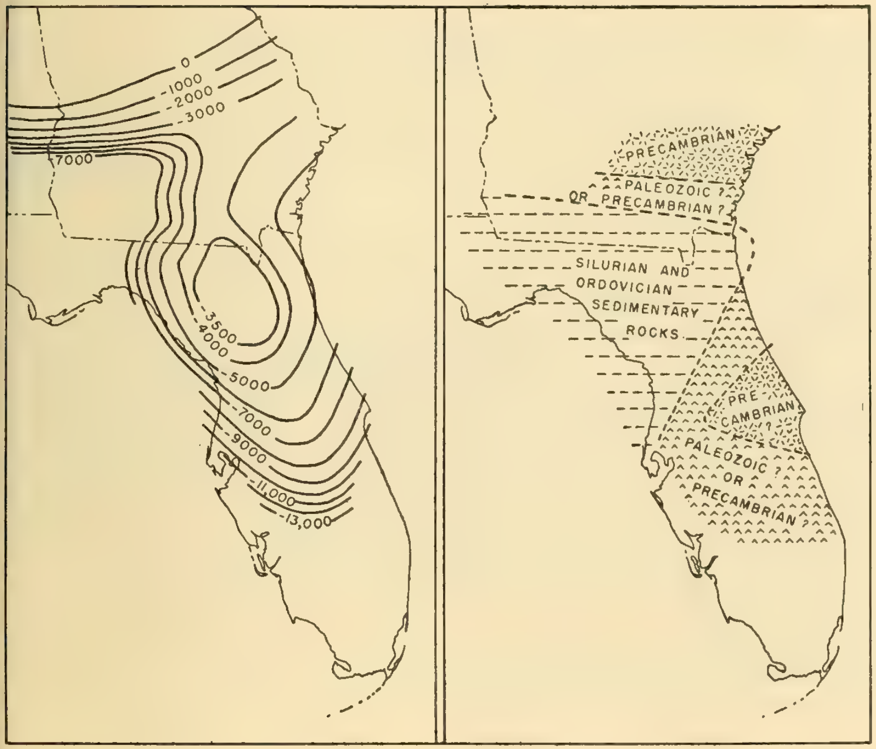


Fig. 41.14. Configuration of surface of pre-Mesozoic rocks in Florida and southern Georgia, and distribution of pre-Mesozoic rocks. Precambrian consists of granite, diorite, and metamorphic rocks; Paleozoic (?) and Precambrian (?) consist of rhyolite, tuff, and agglomerate. After Applin, 1951.

ever, suggests a structural high offset to the southeast, with intrusive igneous rocks, probably Pre-cambrian, exposed in the core. These are flanked on the northwest and southwest by volcanic rocks which may be the equivalent of the Unicoi formation (basal Chilhowee) of the southern Appalachians. Then in northern Florida a basin of Ordovician and Silurian sedimentary rocks occurs, fairly flat-lying and unmetamorphosed (Applin, 1951). These undisturbed Paleozoic strata are southeast of the Appalachian orogenic belt, and pose a rather mysterious problem in tectonics and the evolution of the southeastern margin of the continent.

CRUSTAL STRUCTURE OF GULF OF MEXICO

Geophysical Data

Refractive seismic traverses by Ewing *et al.* (1955) and a magnetic intensity survey by Miller and Ewing (1956) serve as the principal evidence for sediment layering and crustal structure under the Gulf of Mexico. The seismic data are given in Fig. 41.15, and the magnetic data have been used in constructing the geologic cross section of the same figure. Another seismic refraction profile by Antoine (1959) across the Colombian basin from western Cuba to Colombia continues the Gulf of Mexico section to South America. Although the two sections are offset from Yucatan to Cuba, the Yucatan–Cuba tectonic element may be visualized as shown in Fig. 41.15, and the effect of a continuous section obtained, which helps in understanding the constitution and history of the great mediterranean region. Cuba and the Caribbean region will be discussed in Chapter 42.

In making the geologic interpretation the rocks indicated by the various seismic velocities are taken as follows. These are generally the ones suggested by geophysicists in previous references on the Atlantic continental shelf and ocean floor, and in the above articles.

1.8–3.7 km/sec	Unconsolidated and semiconsolidated sediments
4.5–5.2 km/sec	Semiconsolidated and consolidated sediments
5.2–5.5 km/sec	Limestone and dolomite
4.5–5.5 km/sec	Extruded porous volcanic rock. Lower values probably indicate porous rock
5.6–6.1 km/sec	Intrusions in volcanic rock
5.8–6.1 km/sec	Crystalline basement of continent
6.5± km/sec	Gabbroic or basaltic subcrust
7.0–7.5 km/sec	Transition layer, mantle to subcrust
8.0–8.3 km/sec	Mantle (periodotite or eclogite)

Shelf of Gulf Coastal Plain

It may be seen in Fig. 41.15 that the wedge of sediments of the Gulf Coastal Plain thickens nearly to the shelf slope where a total thickness

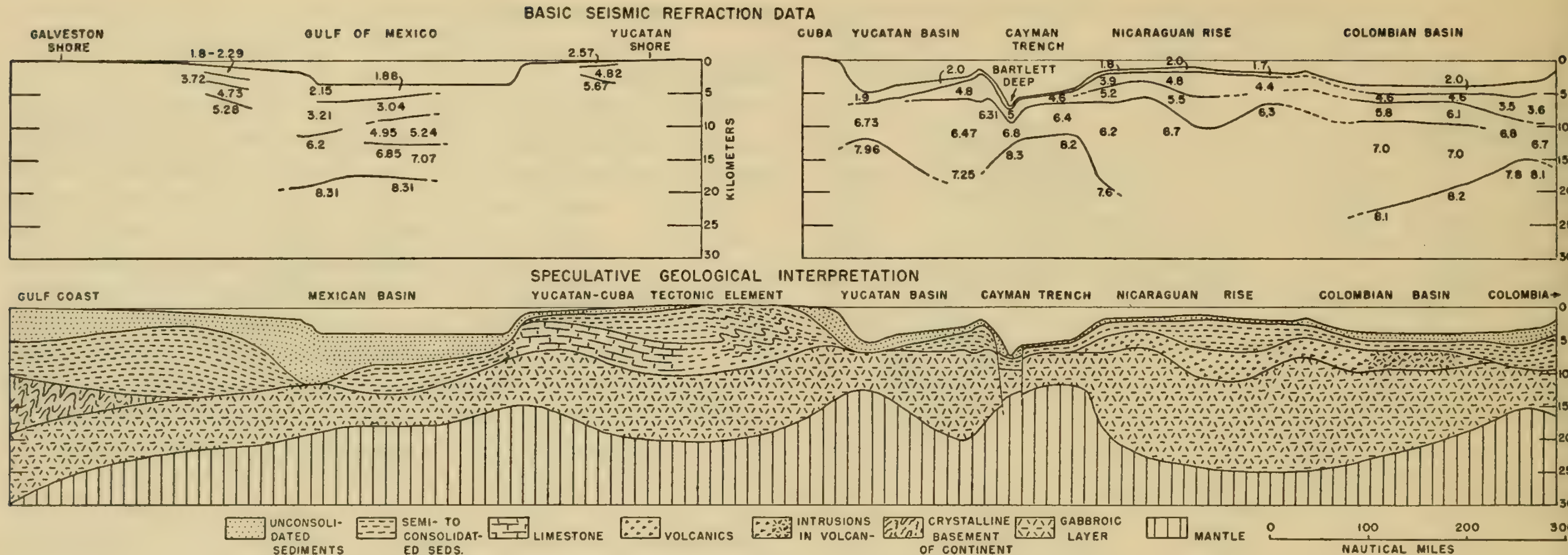


Fig. 41.15. Crustal structure of Gulf of Mexico (Ewing *et al.*, 1955; Miller and Ewing, 1956) and western Caribbean (John Antoine, 1959). The speculative geologic interpretation is slightly altered and somewhat more detailed than given by the authors cited.

of about 45,000 feet of combined consolidated and unconsolidated sediments appears to exist. Refractions from the base of the "consolidated sediment" layer could not be obtained, and it is inferred by Ewing *et al.* (1955) that either a limestone or salt layer of about 5.28 kilometers per second velocity overlies lower velocity sedimentary rocks. The great thickness of the "consolidated" layer may be the result of consolidated carbonate facies, and the boundary shown on the geologic section of Fig. 41.15 may therefore not be a systemic or time boundary. This seems a more logical interpretation than one involving vertical movements of the ocean floor. Offshore carbonate deposition in the form of barrier reefs could have affected the semi-isolation of extensive lagoonal seas for the precipitation of salt and gypsum. Some such retaining form or structure is necessary to produce the Jurassic or Permian evaporite conditions of the Gulf Coastal Plain.

It will also be observed that the sediments near the shelf slope rest directly on the gabbroic subcrustal layer, and that the continental crystalline basement layer does not make an appearance until about the shoreline. This arrangement is concluded by Miller and Ewing (1956) to exist because the magnetic intensity field is remarkably uniform and without conspicuous anomalies from the basin across the shelf slope onto the shelf.

The shelf slope has been considered to be a fault scarp and in addition to indicate that the Mexican basin is a down-faulted depression (Gealy, 1953; Jordan, 1951; Eardley, 1954). The uniform magnetic field

across the steep slope argues against the fault theory, as does also the lack of seismic activity there (Miller and Ewing, 1956).

Mexican Basin

The seismic data indicate that some 30,000 feet of unconsolidated and consolidated sediment under the Mexican basin rests directly on a gabbroic subcrust which in turn is about 25,000 feet (8 kilometers) thick. This condition indicates that the Gulf of Mexico crust is of the oceanic type, but that sediments have been accumulating in the large amounts characteristic of continental borderlands on the gabbroic layer from at least the beginning of Mesozoic time.

Yucatan Platform

The north side of the Yucatan peninsula or platform, the Campeche Bank, is believed to be underlain by limestone or dolomite with only a thin veneer of unconsolidated sediments. The velocity of 5.6 (Fig. 41.15) is regarded by Miller and Ewing (1956) to indicate limestone, dolomite, or crystalline basement, but the exposed geology suggests the presence of carbonates rather than a crystalline basement. The carbonates are believed to be sufficiently lithified and strong to hold up an exceedingly steep slope, which in turn is interpreted to be an escarpment built up by sedimentary processes and not a fault scarp. The uniform magnetic field over the escarpment points to the sedimentary origin.

42.

ANTILLEAN–CARIBBEAN REGION

GEOGRAPHIC PROVINCES

The West Indies were discovered by Columbus when he came ashore on the island of San Salvador. The name Antilles, which comes from the mythical island of Antilia or Antillia, and this in turn possibly from Atlantis, Plato's vanished land in the Atlantic, came to be applied to the islands of the region (Schuchert, 1935). Following the general pattern of use today, the term Greater Antilles will refer to the major islands, Cuba, Jamaica, Hispaniola (the Dominican Republic and Haiti), Puerto Rico (Porto Rico), the Virgin Islands, and the Bahama Islands. See map, Fig. 42.1. Puerto Rico and the Virgin Islands are separated from the volcanic islands on the south by the Anegada Passage, which has been taken by Schuchert (1935) to mark the eastern limit of the Greater Antilles. The Anegada Passage is the site of a submarine channel across the Caribbean submarine ridge, and its shallowest course is over 3000 feet deep. The arc of volcanic islands south of Anegada Passage is known variously as the Caribbees, the Windward Islands, and the Lesser Antilles.

The Caribbean Sea, according to most maps, includes all water south of the Greater Antilles, west of the Lesser Antilles, north of Colombia and Venezuela, and east of Central America. The major basin is south of shallow banks that stretch from Honduras and Nicaragua to Jamaica, and from Jamaica to Hispaniola. It is divided into a western half, the Colombian basin, and an eastern half, the Venezuelan basin, by the Beata ridge which extends southwesterly from Hispaniola. The Tanner basin or deep in the eastern half has a greatest known depth of 16,800 feet. The Aves swell, marked on the north by Aves Island, separates the Venezuelan basin from the Grenada basin, which is bounded on the east by the Caribbees and their supporting ridge.

North of the Rosaline and Pedro Banks and Jamaica, and south of the Misteriosa Bank, the Caymans, and eastern Cuba is a deep, east-west-trending basin with greatest known depth of 22,788 feet. The major basin is called the Cayman trench, and the deep inner trough, the Bartlett. See Fig. 41.15.

GREATER ANTILLES

Cuba

Physiography. Cuba is the westernmost island of the Greater Antilles. It is 100 miles south of Florida, is 750 miles long, and has an average width of 50 miles. The shape of Cuba as defined by the existing shorelines would be considerably changed if the water level dropped only 50 feet. The Isle of Pines and numerous cayos on the north and south coasts would become part of the mainland, and the area would be increased 30 percent (Palmer, 1942). Beyond the 50-foot isobath, deep water sets in almost everywhere.

The principal geomorphic divisions are shown in the upper map of Fig. 42.2.

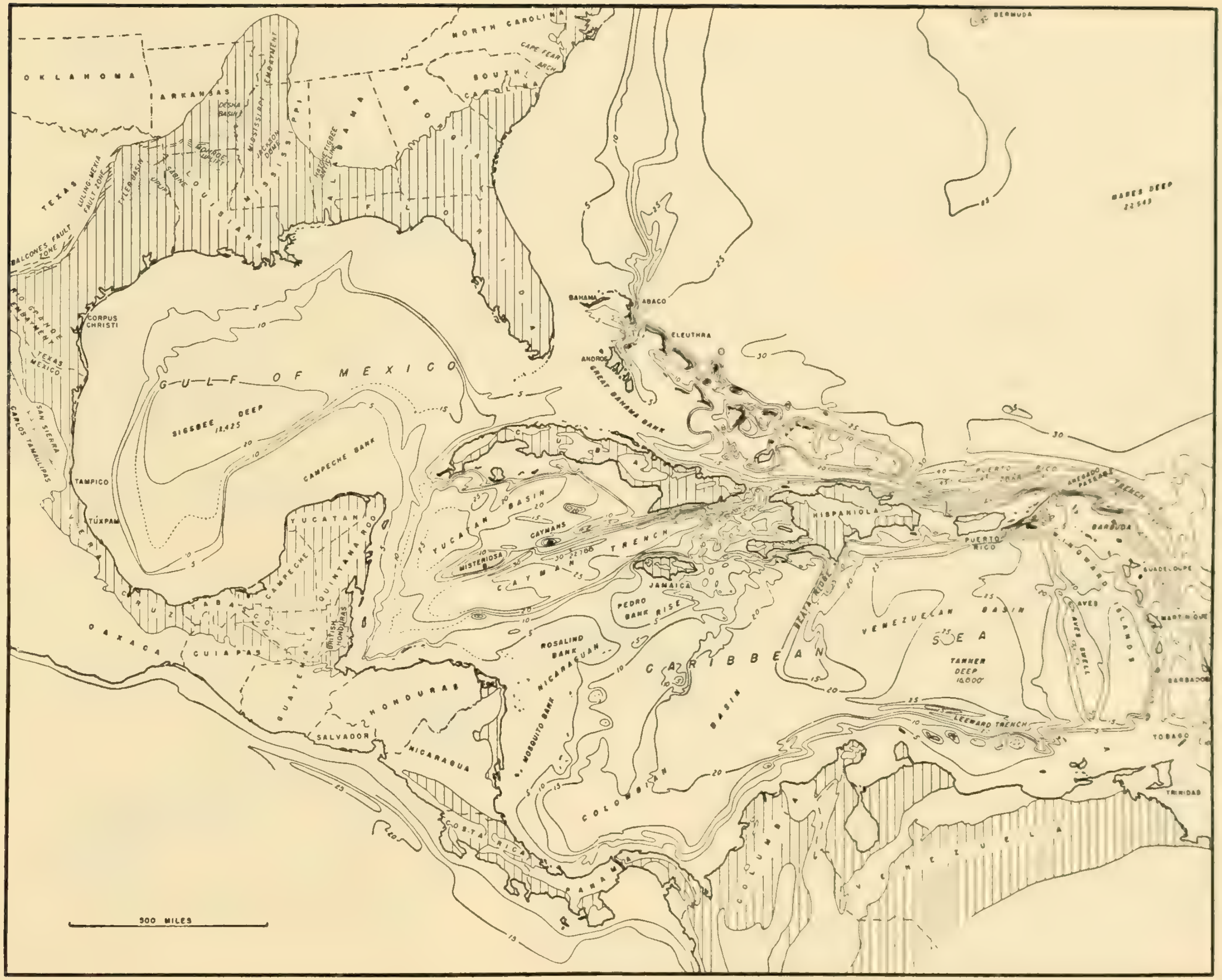


Fig. 42.1. Map of the Gulf of Mexico and the Caribbean Sea regions. The lined areas are underlain by Tertiary sedimentary rocks. The sea and ocean floors are contoured in hundreds of fathoms.

Fig 42.2. Geographical and geological maps of Cuba. Geology after Butterlin, 1956. Facies lines from C. W. Hatten (personal communication) and Wassall (1957) apply to Jurassic and Cretaceous strata.

			CUBA	HISPANIOLA	PUERTO RICO	VIRGIN ISLANDS	JAMAICA
CENOZOIC	RECENT		Calcareous reefs, alluvium	Alluvium	Reefs and alluvium	Alluvium	Reefs of several islands
	PLEISTOCENE		JAIMANITAS reef ls.(20m.)	Reef ls., nepheline basalt, alluvium	SAN JUAN aeolian calcareous ss.	Sand and reefs	[illegible], clay, sand gravel (300m.)
	PLIOCENE		MATANZAS ls. and marl	HINCHE alluvium RIVIERE GAUCHE molasse (500m.)	?	?	[illegible], marl and white ls
	MIOCENE	UPPER	EL ABRA clay & sand (50m.)	ARTIBONITE GROUP: MORNE DELMAS basalt (400m.); LAS CAHOBAS congl., ls., gravel, clay; MAISSADE clay, lignite	GUANAJIBO(?) sandy ls., clay, silt, sand	KINGSHILL marl (180m.)	[illegible] gravels and marles
		MIDDLE	CANIMAR marl & argillaceous ls.(45m.)				[illegible] lavas ?
		LOWER	GUINES ls. (40m.) PASO marly ls.		d'AYMAMON ls. (325m.) d'AGUADA ls. (75m.)		WHITE LS (600m.): [illegible], yellow ls.; [illegible] ls.; [illegible] ls
	OLIGOCENE	UPPER	COJIMAR marl (35m.) REAL congl.	THOMONDE clay and ss.(750m.)	RIO GUATEMALA GR.: CIBAO marl (230m.); GUAJATACA detritus (120m.); LARES ls. (400m.); SAN SEBASTIAN sh., sand, gravl.(300m.)	JEALOUSY GROUP: clay, gray-green (300m.); conglomerate & gray clay (30m.); clay, gray and ls. (90m.+)	MONTPELIER ls ? BROWN'S TOWN
		MIDDLE	JARUCO, marl, congl., sand	MADAME JOIE sh.&ls. (720m.) LA CRETE ss.& ls.			
		LOWER	TINGUARO marl	LIMESTONE	CANAS ls., siliceous sh. (660m.)		
	EOCENE	UPPER	CONSUELO marl JABACO marl & congl. JICOTEA marl	ENNERY ls. (1,000m.) Dolerite & Basalt	COAMO SPRINGS ls., tuffs (300m.)	?	SOMERSET ls. GIBRALTAR ls. SWANSICK ls. SWANSICK ls. TROY dol. & ls.
		MIDDLE	LOMA CANDELA, ls., marl, ss., congl.	PLAISANCE ls. (1,000m.) PERODIN tuffs, sh., & lavas (1,000m.)	RIO JUEYES sh., congl., ls., tuffs (1050m.)	?	?
		LOWER	UNIVERSIDAD marl (130m.) TOLEDO clay CAPDEVILA sh. & ss. LUCERO member, ss., congl.	ABUILLOT sh., ss., congl. (1,000m.)	COROZAL ls. Granodiorite Serpentinized peridotite, cal.tuffs	? Granodiorite	YELLOW LS. (150m.) Porphyry sill (300m.) WAGWATER GROUP, congl., ss., & sh. (460m.) HALBERSTADT GROUP volcanics
		PALEOCENE	MADRUGA ss., clay gravel, congl. (600m.) REMEDIOS crystalline ls.	MARIGOT, congl., ss., sh. (600m.)	Tuffs, breccias, agglomerate, sh., ls., andesitic and basaltic (3000m.)	MOUNT EAGLE volcanics (9000m.) metamorphic rocks, ls., Rudistid ls., andesitic lavas	Grandiorite SEDIMENTARY SERIES, sh. & ls. (2100 to 2400m.) [illegible] SERIES andesites & basalts SUNDERLAND SERIES vol. breccias & tuffs
MESOZOIC	CRETACEOUS	UPPER	northern facies - ls. HABANA southern facies - tuffs	Quartz diorite, serpentinized peridotites, dolerites MACAYA marl & radiolarite (2,000m.)			
		LOWER	Quartz diorite, serpentinized peridotite and gabbro intrusions TUFF SERIES, tuffs, ls., marl. (8,000m)	BASAL COMPLEX, tuffs, andesite, basalt, mica, chlorite, and calcareous schists, amphibole	BARRANQUITAS-CAYEY sh., ls., tuffs (900m.) RIO DE LA PLATA, tuffs, congl., sh., ls. (600m.) ? CABO ROJO rhyolite, granodiorite, granite		BASAL COMPLEX, schist, marble, amphiloolite
	UPPER JURASSIC		VINALES ls. JAGUA schistose ls. AZUCAR ls.				
	MIDDLE JURASSIC		SANCAYETANO sh., ss., slate, phyllite. Includes basal metamorphic complex				

Fig. 42.3. Stratigraphy of the Greater Antilles. After Butterlin (1956) with modification of the Cuban Jurassic sequence. See modifications in text of sections in Hispaniola and Puerto Rico.

Stratigraphy. The succession of rock units of Cuba is given in the chart of Fig. 42.3. The oldest rocks are a metamorphic complex which crops out in the eastern Oriente Province, in two places in central Cuba, and on the Isla de Pinos. It is included in the middle Jurassic by Butterlin (1956) but may be older (Taber, 1934). The complex in the Trinidad Mountains consists of limestones and dolomites and a carbonaceous, chloritic, mica schist. Quartz-garnet-mica schist and epidote and talc schists are also noted (Hill, 1959). No fossils were found. Serpentines of two types occur in the complex, a nodular one derived from periodotite and a fine-grained one derived from microgabbro. The one derived from microgabbro is older and has been affected by two movements, one preserpentinization and one postserpentinization. The rocks are isoclinally folded.

The Jurassic and Lower Cretaceous strata are irregularly treated in the literature, but now it is believed that a continuous sequence exists from the Middle Jurassic to the Tertiary. According to C. W. Hatten of Standard Oil of California (personal communication) the northern succession from the Vinales limestone through the Cretaceous is a carbonate facies closely related to the Florida deposits, and the southern facies is a graywacke-volcanic succession. The graywacke-volcanic facies is called the clastic-volcanic facies by Wassall (1957) and the carbonate facies the limestone and clastic-volcanic facies. North of the carbonate facies is the exaporite facies. These facies hold for the Upper Jurassic and Lower and Upper Cretaceous; the sites of deposition did not shift appreciably during the entire time. The approximate facies zones are indicated on the geologic map of Fig. 42.2. The San Cayetano formation consists of some 35,000 feet of highly folded shales, slates, phyllites, and minor amounts of schist. The Vinales limestone consists mainly of dark gray to black, fairly thin-bedded limestone, but it includes considerable amounts of dark shale and chert. Its thickness from place to place has been variously estimated from 1000 to 5000 feet. It crops out chiefly in parts of western Cuba.

The lower Tertiary deposits are mostly clastic and contain coarse con-

glomerates and sandstones. They have been involved in strong orogeny along with the Cretaceous strata and are separated from the middle and upper Eocene by a major unconformity.

Middle and upper Eocene are found in all the provinces of the island. In this series occur fine conglomerates, sandstones, limestones, marls, and chalks. The Eocene deposits indicate a progressive deepening of the depositional area.

The Oligocene is very well represented. At least seven horizons have been recognized in various parts of the island ranging from the lower Oligocene to an Oligo-Miocene transitional one. It carries a large and well-preserved fauna. The formation is predominantly lime in various stages of induration, and coral reefs are common. The lowest known Oligocene member is a marly shale. Both Eocene and Oligocene contain shale and some marl members that would afford admirable cap rock for petroleum reservoirs.

There was a continuity of deposition from the Oligocene into the lower Miocene. During this period Cuba was submerged except for a few islands. The general aspect was probably not greatly different from that of the Lesser Antilles today, that is, a series of small islands. The deposition of this period is predominantly a hard limestone which has been named the Guines. This limestone forms an interrupted collar nearly around the island as far east as Camagüey, and crosses the island in two low, flat saddles, one in Matanzas and the other in western Camagüey Province. It lies unconformably upon almost all the preceeding formations. Except where folding has subjected it to erosion, the Guines limestone masks the older formations. Geological data are here dependent on geophysics and core drilling.

Deposits of mid- and late Miocene age are limited to a few estuaries that were inundated at the time. They are best developed around Matanzas Bay, Santiago de Cuba, and Manzanillo, and extend but a short distance inland from the coast. The remaining Tertiary deposits are but small local patches along the coast.

The Pleistocene record is confined to well-developed terraces in several parts of the island and to a few scattered unimportant deposits along the coast.

The Upper Cretaceous and the Tertiary, except for the lower Eocene, carry large and well-preserved faunas. These consist of Foraminifera, Radiolaria, corals, echinoids, and mollusks. A noteworthy feature of the Cuban fossil faunas, of both the Cretaceous and Tertiary, is that they are definitely not North American. They are tropical faunas and form a part of a Caribbean unit. This unit is in turn a part of the Mediterranean or Tethyan fauna of the Old World. The Aptychus beds are a striking illustration. Deposits with the same fauna, of the same lithologic aspect, attributed to the same age are found in the Cape Verde Islands and in Persia. Another equally striking illustration is that what appear to be the same species of echinoid occur in the Eocene of both Cuba and Egypt.

Igneous Rocks. Both intrusive and extrusive rocks occur in Cuba. The intrusives are both acid and basic. The acid rocks for the most part occur in the southern half of the island. This is illustrated by a large granite intrusion that borders the Trinidad Mountains on the north and by the granite and other acid intrusions on the southern slope of the Sierra Maestra. These intrusions are relatively not extensive.

In contrast, the basic intrusions for the most part lie in the northern half of the island and are by far the more prominent type. Most of the basic rocks are serpentine. There is no agreement on the age of these intrusions. Most of them occur in the Cretaceous terrane and appear to be post-Cretaceous. Two are known in an Oligocene terrane and appear to cut the limestone of that age. The very extensive intrusions of serpentine and associated rocks in Santa Clara Province are thought to have accompanied the post-middle Eocene period of overthrusting.

Lower Cretaceous volcanic activity was considerable. This is evidenced by thick series of tuffs, volcanic breccia, and flows. At least 6000 feet accumulated in the southern part of Habana Province.

Except in Oriente Province, there is but little effusive volcanic material in the Tertiary. In that province the middle Eocene deposits are largely basaltic. The Sierra Maestra is composed of rocks of this material. Taber (1934) estimates the thickness "to be over 4500 meters and possibly as much as 6000 meters."

The upper Eocene in Oriente Province is also basaltic in part, but much less so than the middle Eocene. In Matanzas Province there are thin beds

of pumice in the upper Eocene. In Camagüey and Oriente there are a few Late Tertiary or Pleistocene flows. Elsewhere the Tertiary is free of volcanic material.

Tectonic History. The Jurassic and Cretaceous tectonic history of Cuba has been interpreted variously by different writers, but this is most probably due to the fact that until recently the facies relationship of the several formations has not been entertained. Unconformities and several pre-Tertiary deformations have been postulated. According to Wassall (1957) the main deformation occurred near the close of Cretaceous time when the southern clastic-volcanic sequence was thrust northward over the carbonate facies and even over the southern margin of the evaporite sequence. Later normal faulting parallel with the facies zones resulted in the dropping of blocks of the thrust sheet into graben. The upfaulted blocks of the thrust sheet were eroded away but the downfaulted masses were preserved, and appear to be out of place in the northern facies unless thrusting of great magnitude is postulated. The age of the graben faulting is not given by Wassall, and it is not known how it fits in the Tertiary chronology of Butterlin, given below.

The serpentine is believed to be tabular, associated with layered gabbro, and carried northward in the thrusting. Others have proposed that the basic plutons were intruded at about the same time as the acidic plutons. Butterlin (1956) suggests that the intrusions occurred near the close of Early Cretaceous time, but evidently Wassall considers the acidic plutons much later than the basic.

Early Eocene time saw much flooding and probably the development of deep water in the west. Close to the mountains conglomerates, sandstones, and shales accumulated, but at a distance marls.

Orogeny then occurred at the close of the early Eocene, probably continuing into mid-Eocene (Butterlin, 1956). The effects are most con-

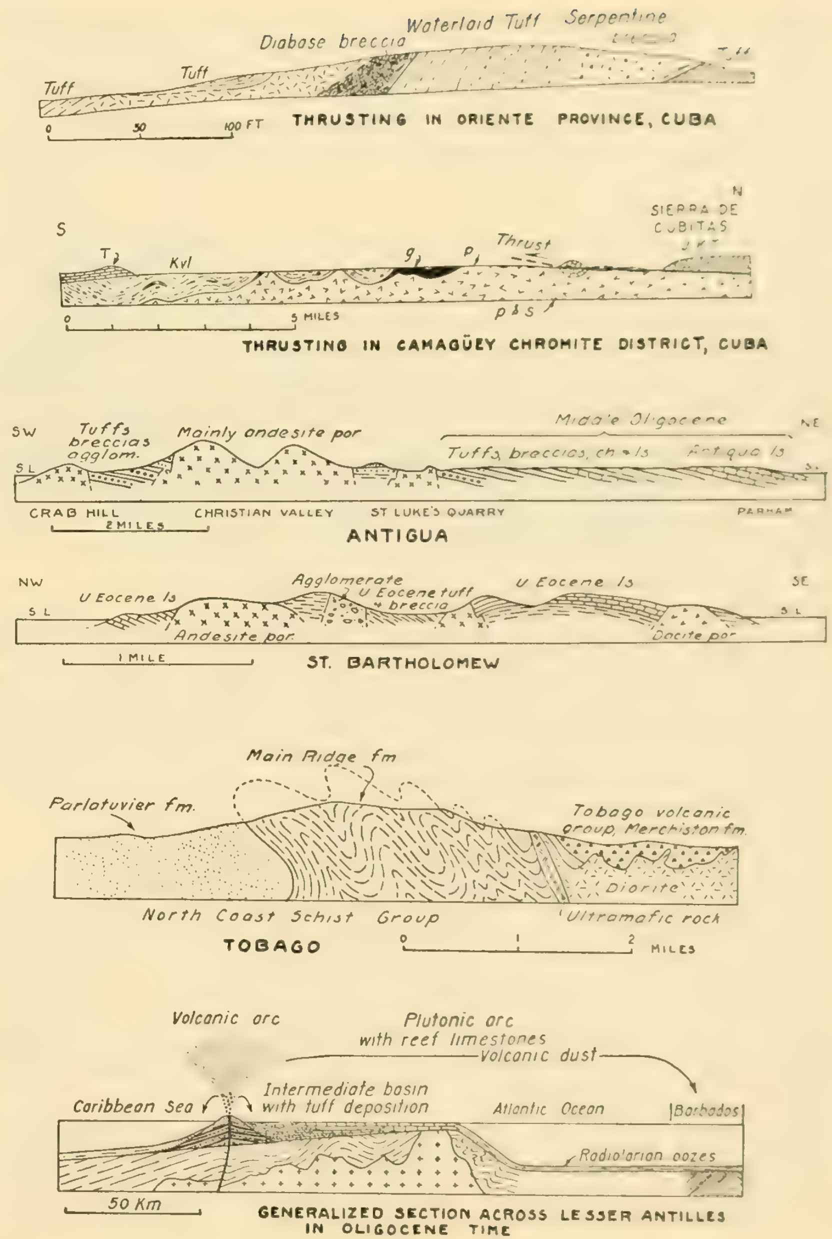


Fig. 42.4. Cross sections of the Greater and Lesser Antilles. The upper two sections are in Cuba, after Thayer and Guild, 1947. The third and fourth sections are across Antigua and St. Bartholomew islands, after Christman, 1950. The fifth section is across Tobago, after Maxwell, 1948. The schists, volcanics, and intrusive rocks are regarded by him as Cretaceous. Lowest section is a hypothetical interpretation by Senn (1940), across the arc of the Lesser Antilles showing the conditions of sedimentation during the Oligocene epoch. The north to south thrusting in Camaguey is now doubted.

spicuous in the western and central provinces. Overthrusting is described in several places along the northern margin of the island (Fig. 42.4) and presumably occurred at this time. All thrusting is now believed to be toward the north. It is not known whether or not this mid-Eocene thrusting of Butterlin is the same as the very Late Cretaceous thrusting of Wassall.

Orogenic movements then probably spread to the southern part of the Province of Oriente. . . . It should be added the upper Eocene begins with conglomerates. Recurrence of volcanic activity is shown by the presence of tuffs and of subsidence basalt dykes, thereupon marls and limestones (San Luis formation) forming a deposit. In the Guantanamo Basin are found thick shales of the same period (Guantanamo formation). At the end of the middle Eocene, in the central and western areas, the sea appears again depositing at first conglomerates and sandstones and, afterwards, limestones and marls (Loma Candela formation). During the upper Eocene, the sea still progresses and if littoral series (conglomerates and marls of Jabaco formation) are found, deep deposits prevail (marls of the Jicotea group of the Jabaco formation and pelagic marls of the Consuelo formation (Butterlin, 1956).

Hispaniola

Physiography. Haiti and the Dominican Republic make up the island of Hispaniola, which is the second largest of the Antilles. It is about 400 miles long and in its widest part 160 miles. The greater part of the island is ruggedly mountainous, with three or more of the clearly defined northern ranges trending N 60° W. The axial or Sierra Central reaches an altitude of 10,249 feet. This is the highest peak in the West Indies.

North of Hispaniola is a narrow submarine trough with a general depth of 12,000 feet, and beyond this is the shallow platform of the Bahama Islands. See Fig. 42.1. Eastward the trough leads into still deeper water, the Puerto Rico trench. South of Hispaniola the narrow shelf soon drops off into the deep water of the Caribbean Sea. Cape Beata is a southward projecting peninsula which continues as a submarine relief feature, the Beata ridge, into the Caribbean basin, and divides it fairly well into eastern and western parts.

The intermontane valleys are thought by some to be of fault origin. This is especially true of the Cul de Sac and the Basin of Enriquillo.

Stratigraphy and Structure. The oldest rocks of Hispaniola are metamorphic and igneous rocks which according to the present literature make up the axis of the Cordillera Central and a large part of the northeastern peninsula of Samana. See Fig. 42.5. Greenstones and amphibolites also occur in the northern part of the island and may be a part of the ancient complex. The quartz diorite is said to be of batholithic proportions.

A new study of the complex has been made by Carl Bowin and he reports on it in a letter to the writer as follows:

Metamorphic rocks occur in central Dominican Republic at the eastern end of the Cordillera Central and continue westward along the northern flank of the Cordillera Central. These metamorphic rocks are probably of early Lower Cretaceous or pre-Cretaceous age although direct evidence as yet only proves a pre-Tertiary age. Thus in central Dominican Republic the oldest rocks do not form the core of the Cordillera Central (as would be concluded from previous reports), but flank the high mountains on the east and north. Towards Haiti, however, the metamorphic rocks may make up the high mountains of the Cordillera Central.

Schistose limestone and quartz-calcite-chlorite-muscovite schists of unknown age are found on Samana Peninsula. The foliation in these metamorphics is reported to trend east-west. Metamorphic rocks are known in the basement rocks that crop out near Puerto Plata on the north coast. However, the lithologies present and their relations are but poorly known. Pre-Tertiary (?) argillites are reported to occur on the south flank of the Cordillera Central and on the south slope of the Cordillera Septentrional, but the grade of metamorphism represented, if any, is unknown.

A large serpentinized periodotite mass occurs in the metamorphic belt in the central part of the country. The intrusive extends northwestward from north of Ciudad Trujillo for a distance of 95 kilometers. A few small peridotite masses are found in the metamorphics along the north flank of the Cordillera Central. These appear to be the westward continuation of the large peridotite mass in central Dominican Republic. Another serpentinized peridotite intrusive, trending N 75° W, has been traced for 80 kilometers along the north coast. A few small bodies of peridotite occur in the eastern portion of the island.

The most detailed work on the pre-Tertiary rocks of the island of Hispaniola has been carried out in central Dominican Republic. Here the metamorphic belt trends NW–SE and consists of primarily epidote amphibolite and schistose siricitic quartz keratophyre. The epidote ampribolites are intruded by several plutons of leucocratic muscovite tonalite and two plutons of gabbro. Both igneous types are probably of early Lower Cretaceous or pre-Cretaceous age.

The amphibolitic rocks are in fault contact with Upper Cretaceous (Cenomanian to Maestrichtian) unmetamorphosed volcanic rocks to the west. These Upper Cretaceous volcanic rocks make up the high mountains of the eastern

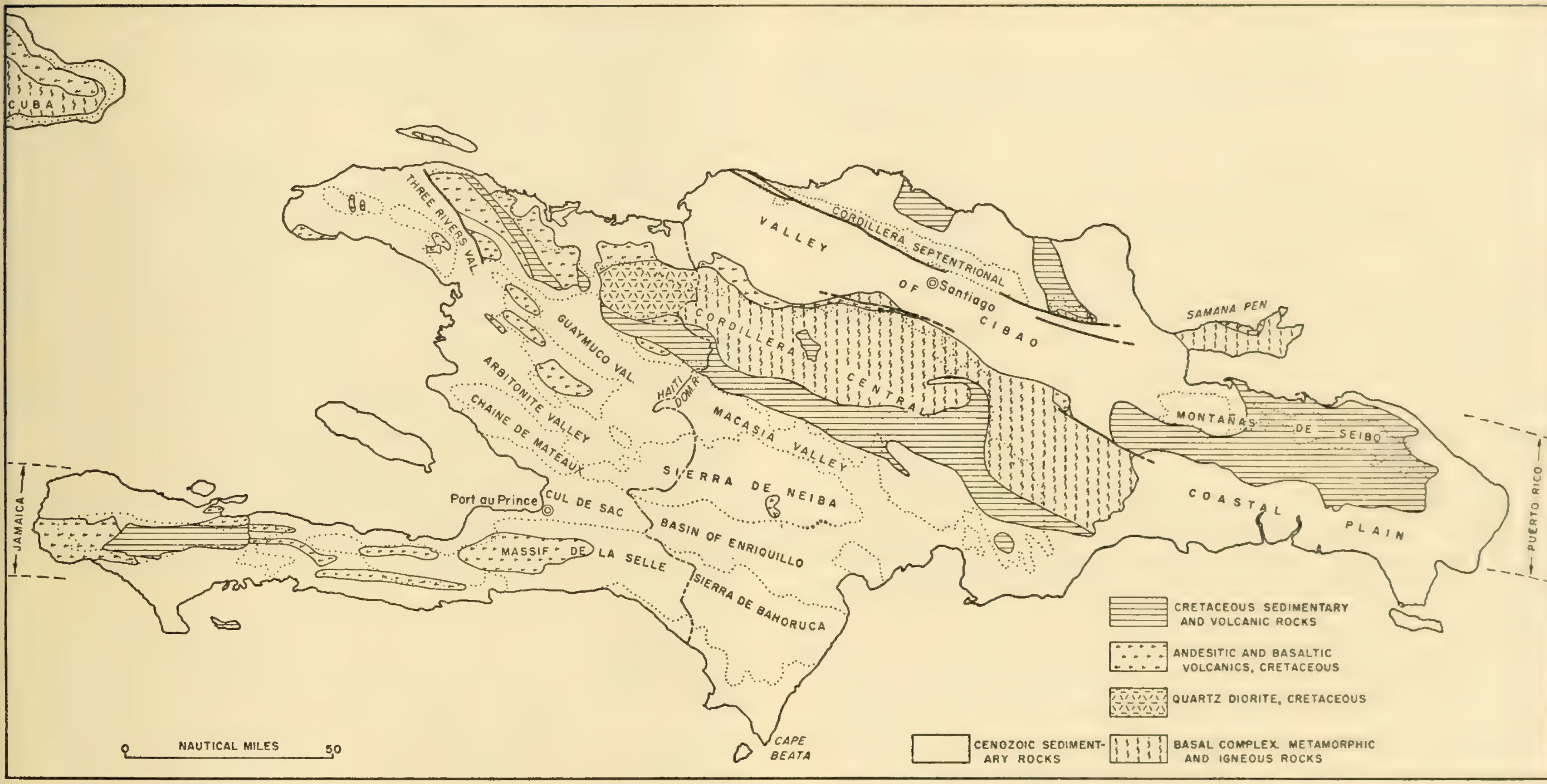


Fig. 42.5. Simplified geologic and terrane map of Hispaniola (Haiti on west and Dominican Republic on east). Geology after Butterlin (1956) and Bowin (unpublished). Terrane from USAF Aeronautical charts. Cretaceous rocks considered mostly Upper Cretaceous by Bowin.

Cordillera Central. They are intruded by hornblende tonalite plutons, at least one of which is of batholithic dimensions. Several plutons and batholiths of hornblende tonalite intrude the metamorphics along the northern flank of the Cordillera Central. The hornblende tonalites are probably all of one general period of intrusion. Cobbles of hornblende tonalite are found in uppermost Upper Eocene conglomerate a short distance north of the Cordillera Central. Thus the hornblende tonalites are considered to be of post-Campanian, pre-Oligocene age. They are probably related to the strong late Eocene deformation that thrust the metamorphic belt northeastward over unmetamorphosed, dated, Lower Cretaceous to Middle Eocene, volcanic and limestone rocks.

The unmetamorphosed rocks lying to the northeast of the overthrust metamorphic belt in central Dominican Republic consists of Lower Cretaceous volcanic rocks overlain by Lower Cretaceous limestone. The Lower Cretaceous section is unconformably overlain by Upper Cretaceous limestone followed

apparently conformably by Paleocene, and Lower and Middle Eocene tuff with lenses of algal limestone.

The eastern part of the Dominican Republic is composed predominantly of fine-grained tuff and interbedded dark gray limestone. These rocks are as yet undated, but are probably of Upper Cretaceous age. Upper Cretaceous sediments are reported from a few localities on both the north and south flanks of the Cordillera Central.

North and south of the Cordillera Central Eocene sections are dominated by limestone. However, in the Cordillera Central and east of it, clastic sediments and tuffs were deposited in earliest Tertiary. Thus in the earliest Tertiary there was a zone of volcanism and uplift in central Dominican Republic. This zone may trend WNW into Haiti parallel to the trend of the Cordillera Central.

The Oligocene and younger sections of the Dominican Republic are dominated by clastic sediments and reflect a complex history of uplift and basin development.

According to Butterlin (1956) the Lower Cretaceous Tuff series of Cuba spreads eastward through Hispaniola, especially in the northern and central regions where andesitic tuffs, basalts, and andesites accumulated.

In the peninsula of southern Haiti thick pelagic limestones accumulated whose fauna bespeaks a Senonian age. This is the Macaya formation. Thick and widespread underwater basalt flows occurred just before and after the limestone depositing epoch, and probably extended westward to Jamaica.

According to Butterlin again, sea flooding and deposition of sediments were resumed in Paleocene time and then lasted until mid-Eocene. Conglomerates, sandy shales, calcareous sandstones, and clastic limestones, the Marigot formation, started the sequence, but these give way in places during early and mid-Eocene time to chalky limestones. In the northwestern peninsula a trough spread to Cuba, and in it thick basaltic and andesitic tuffs accumulated which alternate with thin calcareous layers (Perodin formation). In other areas crystalline or detrital limestones resembling the yellow limestone of Jamaica were deposited and make up the Plaisance and Hidalgo limestones.

In the northern and north-central regions a new disturbance set in. Folding was accompanied by dolerite and granodiorite intrusions. It is impossible to distinguish the folds of this orogeny from the older ones (Butterlin, 1956).

Limestone deposition continued until late Oligocene when renewed orogenic movements set in to last until the Quaternary. From this time on throughout the Tertiary flysch and molasse type sediments accumulated. Tight folding seems the dominant structure with overturning both north and south (Butterlin, 1956).

Considerable attention has been given the longitudinal valleys or basins between the main Sierras. Some, like Butterlin, favor the view that the mountains continued to rise during the late Tertiary and that a gravity flow type of structure developed toward the basins. Woodring *et al.* (1924) describe the bounding faults as overthrusts. Rich (1956) treats the Cul de Sac as produced by recent upfaulting and upbowing of the bounding mountain block. Bucher (1950) postulates a good deal of strike slip along bounding faults as sympathetic fractures to eastward movement of the great Caribbean block. Hess and Maxwell (1953) show the southern peninsula and the Sierra de Bahoruca to have moved many miles from a west-lying position to its present position, and hence a wrench fault of great magnitude to lie along the south side of the Cul de Sac and the Basin of Enriquillo. Several have related the graben-like depressions to the Cayman trench which projects to the Cul de Sac.

Puerto Rico

Physiography. The island of Puerto Rico is roughly rectangular and is about 35 miles wide and 105 miles long. See Fig. 42.6. Its highest peak is 3750 feet above sea level, whereas the Puerto Rico trench immediately north of the island is 27,972 feet deep. The absolute relief between the two is thus 31,700 feet. The plateau-like ridge upon which Puerto Rico occurs also supports the Virgin Islands to the east. The slope into the trough is in the proportion of one mile vertical to thirteen horizontal. See Fig. 42.1.

To the south of the Puerto Rico and Virgin Island platform the bottom slopes steeply and, within 55 miles, is 17,000 feet deep. This is the site of a submarine trench that leads northeastward to the Anegada Passage. The bottom of the trench is generally 15,000 feet deep but rises to about 3850 feet below sea level at the summit or Passage.

To the west, Puerto Rico is separated from Hispaniola by the Mona

Passage, where the water ranges from 1200 to 3760 feet in depth.

The central part of Puerto Rico is a rugged, east-west-trending mountainous mass of the basement complex rocks, and averages about 2000 feet in height.

A coastal plain is particularly prominent along the north side, and a more limited one occurs along the south side. These have been studied in detail by Zapp *et al.* (1948). A rugged foothills belt flanks the south side of the central Cordillera.

Geology. Kaye (1957) notes two major structural and stratigraphic rock units in Puerto Rico: the older complex, ranging in known age from Late Cretaceous to late Paleocene or early Eocene, and the middle Tertiary sequence, ranging from late Oligocene possibly to late Miocene. The former rocks are eugeosynclinal in character and the latter nonvolcanic, made up dominantly of calcareous marine sediments. The middle Tertiary crops out on the north and south sides of the island and in structural troughs on the west coast. On the north coast the beds dip gently to the north, and, except for slight terracings and a flexure at the northwestern corner of the island, are not folded. The middle Tertiary sequence on the south side of the island is somewhat folded. Seismic-reflection studies of the north coast indicate, however, a pronounced northward thickening, possibly some folding, and unconformities at depth. Unconformities which may be local have also been noted at several places on the surface.

Berryhill *et al.* (1960) have detailed the Upper Cretaceous stratigraphy and facies changes, which may be summarized as follows:

> Rocks of Late Cretaceous age (Turonian to Maestrichitan) in Puerto Rico are of three types: (1) primary volcanic rocks, including tuffs, tuff breccias, and lavas; (2) intermixed pyroclastic and epiclastic rocks, including volcanic conglomerates, volcanic sandstones, and volcanic siltstones; and (3) limestones, most of which were formed as reefs around volcanic islands. These rocks, which have a maximum thickness of more than 20,000 feet, crop out along the crest and flanks of a complexly faulted, northwestward-trending anticlinorium that forms the mountainous core of Puerto Rico.

The major aspect of the structure of the eastern part of the island is anticlinal which Berryhill *et al.* believe is due to doming of the strata during intrusion of a batholith in early Tertiary time. See map, Fig. 42.6.

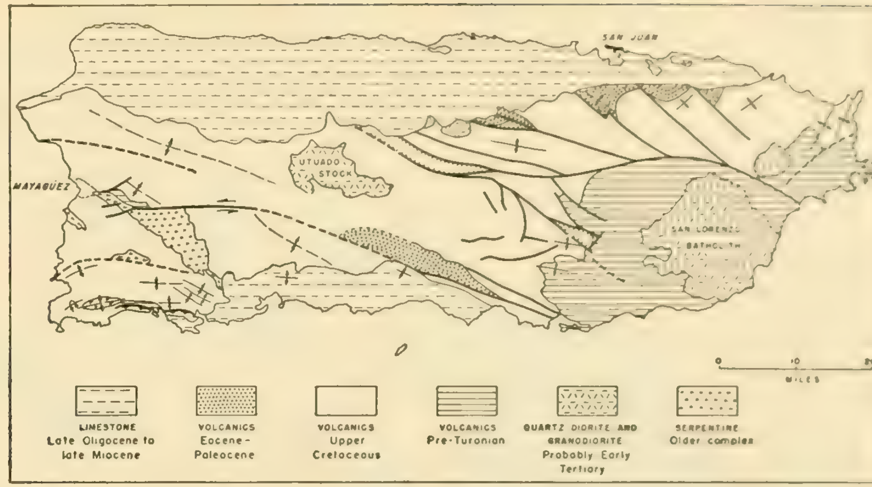


Fig. 42.6. Geologic map of Puerto Rico. Compiled from Kaye (1957), Berryhill, Briggs and Glover (1960), and Mattson (1960). Hachured faults indicate graben.

> Complex faulting that accompanied the batholithic intrusion helped to accentuate the anticlinal structure but in some places modified it. The regional trend of the main faults and also many of the subsidiary faults is west-northwest, but some of the subsidiary faults diverge from that general pattern.
>
> Two faults of regional significance are recognized in eastern Puerto Rico. One crosses the northern part of the island, and the other traverses the south-central part.
>
> Movement along the northern fault appears to have been largely transcurrent. The crustal block north of the fault apparently has moved eastward relative to the block south of the fault.
>
> The subsidiary, northwestward-trending faults on the north formed as tears along the main fault. Movement along most of these subsidiary faults appears to have been both horizontal and vertical because of rotational movement of the blocks formed by the faults. Associated with the northern fault are two grabens. . . . A third, smaller graben, . . . lies south of the northern fault. The stratigraphic displacement at the southeastern end of the largest of these three grabens is approximately 6,000 feet.
>
> The second regional fault, which trends west-northwestward across the southern part of the island, appears to be in part a transcurrent fault and in part a high-angle reverse fault which dips about 70° toward the southwest. The stratigraphic displacement along this fault, based on good stratigraphic control is about 12,000 feet.
>
> The pattern of faulting is related to the crude west-northwestward alignment

of the plutonic intrusive bodies which form a belt across the island. Moreover, the northern major fault roughly coincides with the belt of pillow lavas and volcanic breccias that were extruded during Robles (Late Cretaceous) time. That belt of volcanism may have been located along a regional line of weakness and the younger plutonic intrusives may have moved upward in part along this same general zone.

Although the general anticlinal structure of eastern Puerto Rico is probably a result of doming by a batholith, several localized anticlines and synclines have been formed by the movement of fault blocks (Fig. 42.6). The largest of these secondary structures is the northeastward-plunging, faulted anticline near the northeastern corner of the island. The Luquillo Range is the northwest limb of this breached and faulted fold. This structure probably was formed by compression from the northwest as the crustal block north of the transcurrent fault moved eastward. Tight folding is localized near some of the faults but is not extensively developed in eastern Puerto Rico.

By comparing Kaye's and Berryhill's analysis with that of Butterlin (chart, Fig. 42.3) it will be seen that Butterlin suggests older rocks than they found on the island, and that late Eocene and early Oligocene formations are present whereas they indicate a hiatus for this interval.

Butterlin also points out that broad arching with an east-west axis was the dominant part of the mid-Miocene disturbance.

Berryhill (1959) has elaborated on the transcurrent faulting to the effect that the two principal faults or sets of faults divide the island into three blocks, with the northeastern and southwestern blocks having moved toward the southeast and the central block toward the northwest. This is presumed to reflect eastward movement of the Caribbean block. He assigns the major faults to an Eocene age, whereas Kaye recognized the many "block-faults" as late Pliocene and early Pleistocene.

Isla Mona and the Mona Passage

Isla Mona, 21 square miles in area, and Isla Monito, less than one quarter square mile, are situated in the Mona Passage between Puerto Rico and Hispaniola. Isla Mona is a limestone tableland bounded by steep to vertical cliffs except for a narrow coastal terrace about its southern perimeter (Kaye, 1959). The Isla Mona limestone forms most of the mass of both islands and is probably early or middle Miocene. Dips up to 3½ degrees are visible in the cliffs. In places a thin cavernous limestone, the Lirio, overlies the Isla Mona, and in one place a small angular unconformity is visible. The Lirio is Pliocene or Pleistocene in age.

The great purity of the Isla Mona limestone indicates that it was deposited in an oceanic reef environment far from land, and from this it is deduced that the Mona Passage was in existence in Miocene time (Kaye, 1959).

Jamaica

Physiography. Jamaica measures 144 miles from east to west and has a greatest width of 49 miles. It is very mountainous, with about one-half of its area 1000 feet above sea level and much of it over 2000 feet. The principal range, called the Blue Mountains, occupies an axial position at the east end of the island, and has a sharp crest and numerous, generally cloud-wrapped peaks, the highest of which is 7520 feet above sea level. From the sea on the north, the land rises in gentle hills to the higher country, but on the south high cliffs and abrupt precipices mark the shoreline. See Fig. 42.7.

The relief of Jamaica is of four major types: (1) the interior mountain ranges, constituting the nucleus of the island; (2) an elevated and dissected, arched and karsted, white limestone plateau which surrounds the interior mountains and ends abruptly toward the sea, occupying in all fully four-fifths of the total area; (3) the coastal bluffs or back coast border of the seaward margin of the plateau; and (4) a series of low flat coastal plains between the sea and the back coast border (Schuchert, 1935).

Jamaica is separated from Cuba by 90 miles of water, and the marine basin between is the Cayman trench, here everywhere more than 15,000 feet deep and directly off Cuba, 21,000 feet deep. The long and narrow peninsula of Haiti is about 90 miles northeast of Jamaica, and the two islands are separated by water which has a general depth of over 4000 feet. On the south side of Jamaica lies the Caribbean Sea, whose bottom sinks to 13,800 feet. From the island to Honduras it is 900 miles, and the intervening area is mainly shallow water. It is a broad platform on which the Mosquito, Rosalind, and Pedro banks occur, and which drops off steeply into the Cayman trench on the north and slopes gently into the Caribbean on the south.

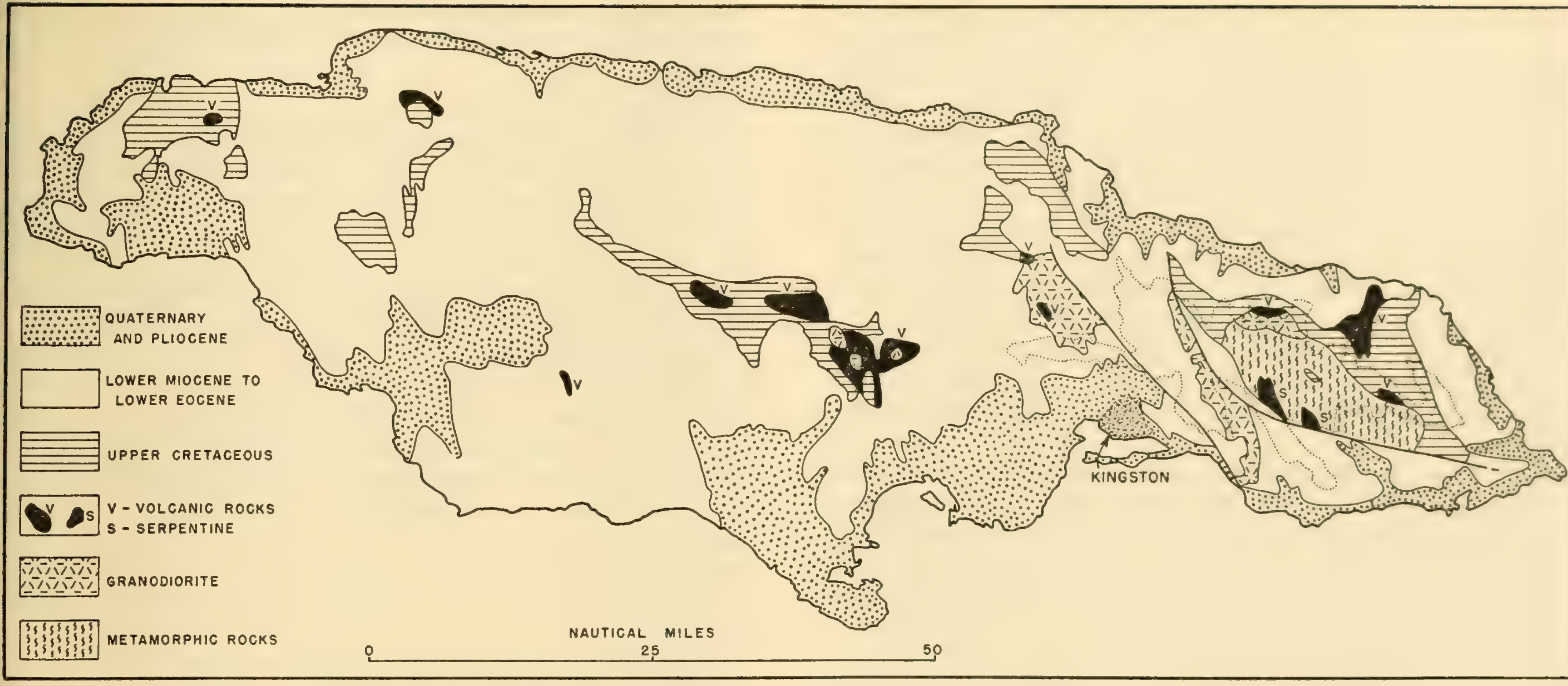


Fig. 42.7. Geologic map of Jamaica. Shoreline and 2000-foot contour (dotted line) from *World Aeronautical Chart No. 647*. Geology adopted from Butterlin, 1956. Dotted contour line outlines Blue Mountains.

Geology. Jamaica, like the other islands of the Greater Antilles, has a basal complex in part older than Late Cretaceous. Three cycles of deposition followed the basal complex, each with a sequence of conglomerate, sand, shale, mudstone, calcareous shale, and limestone, and each separated by an unconformity (Butterlin, 1956). The third cycle ended in early Eocene with intense deformation and intrusions. Thrusting has been noticed and is assigned to this time.

The Yellow limestone of mid-Eocene age was then deposited. White limestone accumulation continued to mid-Miocene time when block-faulting occurred. The faults trend generally north-northwest or north-west, and rejuvenate in places the earlier structures. Extrusions of lavas also occurred.

During the Pliocene, block faulting continued and raised up the calcareous tablelands and in places tilted them. The Cayman Islands across the trench were possibly connected with Jamaica before the block faulting (Butterlin, 1956).

Virgin Islands and Anegada Trough

A bank not more than 165 feet deep extends 100 miles eastward from Puerto Rico like a crescent curving northward. From this bank rise about

100 islands, cays, and rocks which are known as the Virgin Islands. The bank is terminated on the south by the Anegada trough, named from the passage, previously mentioned. Taber (1922) points out that the south side of the trough near the island of St. Croix is a great escarpment which descends 14,130 feet in less than 5 miles, and thus has an average slope of 30 degrees. This he regards as a fault scarp due to vertical movement. As will be related later, Hess believes it is due chiefly to horizontal movement. On the basis of biogeographic data, Schuchert thinks the Virgin Islands were joined to St. Croix across the Anegada trough during either the Miocene or the Pliocene, and that they have been separated due to block faulting along the Anegada Passage in fairly recent times.

It appears that Puerto Rico, the Virgin Islands, and the island of St. Croix developed as a unit, and that the geology of the Virgin Islands, if only partially exposed, fits that of Puerto Rico. See chart, Fig. 42.3.

After the igneous activity and deformation of the older series of Puerto Rico that is also believed to form the foundation of the Virgin Islands bank, a mountainous upland probably existed. According to Meyerhoff (1927), fluvial erosion reduced the mountainous upland to an imperfect peneplain in early Eocene time. The relatively level summits of the upland of Saint John, 1000 feet above sea level, are a remnant of the old surface. Uplift in late Eocene time resulted in dissection of the old surface, and all but the central cores of the present larger islands were reduced to a late mature of old surface about 800 feet below. Only a few remnants of this second surface have been preserved, because a second uplift in early Oligocene time was followed by about 500 feet of downcutting.

The third cycle of erosion formed the lower peneplain of Puerto Rico, as well as the mature to old surface which extends beneath the coastal plain on St. Croix and Vieques, and which underlies remnants of the coastal plain on the submarine platform. Formation of the lower peneplain was followed by subsidence and deposition of coastal plain sediments in the middle Tertiary, and during late Tertiary time uplift exposed the coastal plain marls and limestones to dissection. The Tertiary deposits collected in the entire area now constituting the submarine platforms of the islands. Toward the close of the Tertiary, differential movement or warping caused submergence of the eastern Puerto Rico and Virgin Islands region, while western Puerto Rico remained elevated.

Bahama Islands

Physiography. The Bahama Islands stretch for 900 miles in a northwest-southeast direction in front of southern Florida, Cuba, and Haiti, and include some 29 inhabited islands, 661 keys, and 2387 rocks. The Bahamas are all very low, flat islands and resemble most the coast and keys of southern Florida. All the islands, keys, and rocks rise from a platform that is roughly triangular, with the narrow base of the triangle on the northwest. See map of Fig. 42.1. It is bounded on the west by the Florida Channel, which separates it from Florida by a distance of 50 miles; on the south by the Bahama Channel, which separates it from Cuba by an equal distance; and on the east by the Atlantic Ocean. The greater part of the platform is covered by water only 3 or 4 fathoms deep, but in part it emerges slighty above sea level, forming low islands. Great submarine valleys, such as the Tongue of the Ocean, Exuma Sound, and the Providence Channels, form deep indentations in the platform. On the east, the platform drops off abruptly to oceanic depths (2600 fathoms, 15,600 feet). The extensive shallow banks are remarkable for their white lime oozes. See Fig. 43.4.

Submarine Canyons. The great submarine valleys, which are manifestly a very important character of the Bahama platform, are reviewed as follows by Hess (1933):

(1) The longitudinal valleys have a general NW–SE trend for the greater part of their lengths, but short steep cross valleys at right angles to this trend connect the longitudinal valleys with the ocean.

(2) So far as the information goes, it appears that the valleys slope continuously from the shallowest parts of their upper reaches (720 fathoms, 4,320 feet below sea level) to the floor of the ocean basin proper (2,500 fathoms, 15,000 feet). The longitudinal valleys have gradients of approximately 15 to 20 feet to the mile, and apparently have gently sloping undulating bottoms, from 720 fathoms (4,320 feet) to about 1,000 fathoms (6,000 feet).

(3) The cross valleys have steeper gradients, 100 feet to the mile, from 1,000 fathoms (6,000 feet) to the floor of the ocean at 2,500 fathoms (15,000 feet). They have the typical V-shaped cross profile of a youthful river valley, and some have a distinct inner gorge near the center.

(4) Where examined, the outer rims of all the valleys rise steeply, perhaps even as vertical cliffs, from a depth of about 500 fathoms (3,000 feet) to the edge of the platform.

(5) The Tongue of the Ocean and Exuma Sound Valleys are parallel, and about 50 miles apart, but the Tongue of the Ocean slopes continuously northwest from its shallowest point at a depth of about 720 fathoms (4,320 feet), whereas Exuma Sound, starting from a similar depth, slopes continuously in the opposite direction, southeast.

(6) Where "tributaries" meet the "main stream" they appear to do so at the same level or "at grade," and where the valleys enter the ocean basin proper, they do so at the same level.

Andros Island Deep Test. A deep test well was drilled on Andros Island of the Bahamas to a depth of 14,585 feet, and enhances our knowledge of this little-known region immensely. The following details were given orally by Maria Spencer at the St. Louis meetings of the American Association of Petroleum Geologists, 1949, and taken down as notes by the writer.

Log of Andros Island Deep Test	Depth, Feet
Recent and Miocene (?)	
Limestone as at surface. Corals and bryozoans	0–530
Limestone, dolomitized	530–1625
Coquina	1625–2200
Eocene	
Coquina of microfossils	2200–2640
Alternating limestone and dolomite	2640–4640
Paleocene (?)	
Dolomite, fine-grained	4640–6220
Dolomite and chalky limestone	6630–7590
Paleocene or Upper Cretaceous	
Dolomite, brown	7990–8760
Upper Cretaceous	
Dolomite, tan, granular	8760–9760
Dolomite, coarsely crystalline, cavernous	9760–10,035
Limestone, part brecciated, part chalky, cemented with brown dolomite	10,036–10,660
Cavernous	10,660–10,709
Dolomite, fine-grained	10,709–11,940
Limestone, creamy white, chalky	11,940–12,480
Lower Cretaceous	
Dolomite, crystalline and porous Sunnyland zone in Florida	12,480–13,710
Bottomed	14,587

Spencer commented that the Upper Cretaceous section has the same thickness as that of Florida, but it consists of dolomite and limestone, whereas the Florida section is nearly all limestone. The base of the Lower Cretaceous was not reached in the Bahama test, but the 2100 feet known consists mostly of crystalline dolomite, whereas the Florida section consists of limestone, anhydrite, and dolomite.

Reef Building. Heretofore it could be said only that reef limestones are prominent in many places on the Bahama Islands and have been studied below sea level. A bore hole 395 feet deep on New Providence Island passed through Pleistocene and into Miocene reef material (Hess, 1933). The calcareous material consisted mostly of calcite to a depth of 165 feet, and below it was mostly dolomite. The porosity decreased to 5 percent at the bottom of the hole. Hess recognizes nearly everywhere almost clifflike dropoffs of the submarine canyon walls, ridges, and platforms, from the surface down to a depth of 4000 feet, and believes this feature could not be accounted for by erosion, but on the contrary to reef upbuilding. He finds no geophysical evidence to dispute a conclusion that the reef material may be 4000 feet thick on the Bahamas, and believes it may include the entire Cenozoic section if not also the Upper Cretaceous. His conclusion in theory if not in magnitude proved correct when the deep test described above was drilled. It is concluded that most of the Bahama platform area was a site of subsidence and deposition during the late Jurassic, early Cretaceous, late Cretaceous, and parts of the Cenozoic. The foundations of the Bahamas have been regarded as volcanic by some; but this, in light of present stratigraphic and tectonic data, is only possible below a depth of 15,000 feet.

LESSER ANTILLES

Divisions

The Lesser Antilles, also known as the Caribbees, are an island festoon that extends from the Anegada Passage on the north 430 miles to the Island of Grenada on the south. See maps, Fig. 42.1 and 42.8. Several

Fig. 42.8. The Lesser Antilles showing the Limestone Caribbees (also called older and outer) the Volcanic Caribbees (also called younger and inner).

divisions may be recognized if the submarine features are considered. First, on the west is the Aves Bank that ends on the north at its only emergence, Aves Island. It is slightly convex eastward. The next division is the Grenada basin, a broad and not very deep depression, which bounds the island festoon of the typical Caribbees. These islands are Pliocene–Pleistocene volcanic cones, and they form the inner or younger Caribbees. Outside the volcanic arc and at the north are the Limestone Caribbees, another row of islands which seem to merge with the younger volcanic islands. Outside of the Limestone Caribbees is the deep, narrow Brownson trough, which shallows southward and ends in the wide Tobago trough. East of the Tobago trench is the Trinidad–Barbado element that extends northward in the form of the Barbado submarine ridge outside the Puerto Rico trench.

The volcanic or younger Caribbees begin on the north with Saba (2820 feet high) and extend in succession through St. Enstatius (1950 feet), St. Christopher (4314 feet), Nevis (3596 feet), Redonda (1000 feet), Montserrat (3002 feet), Basse Terre of Guadeloupe (4869 feet), Isles des Saintes (1036 feet), Dominica (4747 feet), Martinique (4428 feet), St. Lucia (3145 feet), St. Vincent (4048 feet), the Grenadines (a series of rocky islands on a narrow bank nearly 100 miles long), and finally Grenada (2749 feet). A considerable number of these islands have well-preserved cones. Some volcanoes are still active, notably La Soufriere on St. Vincent, which erupted violently in 1902 and killed 2000 people, and Mont Pelée on Martinique.

The Limestone Caribbees are characterized by limestones and pyroclastics into which various hypabyssal rocks have been intruded; these are overlain by younger marine limestones. To the Limestone Caribbees belong Sombrero, Anguilla, St. Martin, St. Bartholomew (Barthelémy), Barbuda, Antigua, Grande Terre of Guadeloupe, Desirade, and Marie Galante. Woodring (1928), Senn (1940), and Maxwell (1948) have summarized the geological history of this group.

Outer Limestone Caribbees

The following summary is principally from a report by Maxwell (1948). In the outermost islands of the Limestone Caribbees, Sombrero and

Barbuda, only Quaternary limestone is exposed. Volcanic rocks crop out on Anguilla, St. Martin, and St. Bartholomew. Oligocene limestone rests unconformably on the basement rocks of Anguilla, and Eocene or Oligocene limestone covers the volcanics on St. Martin. The basement of St. Martin consists of well-stratified, strongly folded and metamorphosed tuff, tuff-breccia, and somewhat crystalline limestone, intruded by a quartz diorite-pyroxene diorite complex. According to G. A. F. Molengraaff (1931) the sedimentary material of the basement may be of Cretaceous age, but Christman (1953), as a result of work on St. Martin, St. Bartholomew, and Antigua, states that "there is apparently no Cretaceous basement in the Lesser Antilles." Miocene limestones are nearly horizontal and have been deformed to a less degree than the Oligocene and upper Eocene.

The oldest rocks of St. Bartholomew consist of volcanic debris and an overlying upper Eocene limestone; both are intruded by an andesite porphyry. Also found intruding the Eocene limestone beds are a volcanic agglomerate and a dacite porphyry (Christman, 1953). See cross section of Fig. 42.4.

Antigua and Desirade likewise belong in the outer islands of the Limestone Caribbees. Both have volcanic basement rocks. In Antigua, gently dipping tuffs in the central plain are overlain conformably by the Antigua limestone of middle Oligocene age. The tuffs become coarser to the southwest. At Crab Hill, Christian Valley, and St. Luke's Quarry, intrusive andesite porphyries cut the series (Christman, 1953). See section in Fig. 42.4. Desirade possesses a basement of intrusive granodiorite, with contemporaneous andesite and rhyolite flows. Miocene limestone unconformably overlies the basement.

Grande Terre and Marie Galante are the southernmost, and also the innermost, of the Limestone Caribbees. The latter is covered by a cap of recent limestone (Woodring, 1928). Grande Terre, however, has a basement of granodiorite which is overlain unconformably by lower Miocene tuffs and limestone (Senn, 1940).

To summarize, the outer islands of the Limestone Caribbees are characterized by a basement of lava flows and coarse volcanic debris of late Eocene age, or younger, and andesitic to dioritic rocks of post-late Eocene age which intrude the volcanics. Oligocene and younger beds are mostly limestone, and volcanic debris is fine-grained (as the Central Plain tuff of Antigua), where present. Apparently, these islands were active volcanic centers in late Eocene and Oligocene time and have since received volcanic debris only sporadically and from a distance. They have not been disturbed much by crustal deformation in post-Oligocene time.

Inner Volcanic Arc

The inner arc of the Lesser Antilles, stretching from Saba to Grenada, is characterized by Recent or subrecent volcanic activity. Tuffs of Oligocene age on Martinique (Senn, 1940) and Carriacou (Trechman, 1935) represents the oldest beds identified in the inner arc. Apparently volcanic activity started here about in early Oligocene time and continued with few interruptions to the present. As in the older (pre-Oligocene) volcanics of the Greater and Lesser Antilles, andesites predominate, with basalts and dacites also present. The andesites and basalts of the more recent volcanoes contain hypersthene as a common constituent, whereas the mineral seems to be extremely rare in the pre-Oligocene volcanics. The significance of this mineralogic variation is not apparent.

The volcanoes do not rest on the crest of the swell toward the south; there they are found some 30 to 40 miles west of the crest. At the north end of the island arc, however, they are approximately at the crest, and it happens that here the islands are made up largely of sedimentary rocks.

According to Hess (1938), sonic soundings show a series of peaks on the western flank of Aves swell, parallel to the Lesser Antilles arc and 250 kilometers west of it. Profiles across the peaks strongly suggest submerged volcanoes. The lack of seismic activity along the greater part of the Aves swell in the vicinity of the peaks suggests that if they are volcanoes, they are extinct.

Margarita and the Dutch Leeward Islands

The following is abstracted from Maxwell's report (1948). The northern part of Margarita is composed of paraschists intruded by quartz diorite and serpentinized peridotite. A zone of slightly metamorphosed sediments lies south of the schist area, and unmetamorphosed sediments of

Late Cretaceous age are folded into a syncline along the south coast. Upper Miocene sediments lie unconformably on the lower Tertiary-Cretaceous section. In Margarita, as in Tobago, the foliation in the schist strikes slightly north of east and dips steeply to the southeast. Locally, the peridotite shows relatively low-temperature hydrothermal alteration, probably related to the diorite intrusions, though this is the only evidence bearing on the relative age of the peridotite and diorite. The diorite has suffered strong shearing, as in Tobago. The period of major deformation was post-Middle Cretaceous and pre-Maestrichtian. Detrital grains of chromite and enstatite in Middle Eocene sands prove that the ultramafic mass had been exposed to erosion by that time and hence is pre-Middle Eocene in age. Sometime between the middle Eocene and upper Miocene, a period of moderate deformation folded the Cretaceous and lower Tertiary sediments into a syncline, with its axis approximately parallel to the foliation in the schist.

The Dutch Leeward Islands, Aruba, Curacao, and Bonaire, are volcanic in character, comparable with the Greater Antilles and the outer islands of the Lesser Antilles. A deformed basement of intrusive and extrusive gabbroic and dioritic rocks with intercalated radiolarian cherts, limestone, and graywacke is present on all three islands. Quartz-augite diorite in Curacao (G. J. H. Molengraaff, 1931) and quartz diorite in Aruba (Westermann, 1931) intrude the basement rocks. On Bonaire, limestone of latest Cretaceous age unconformably overlies the basement rocks, which likewise are believed to belong to the Upper Cretaceous (Pijpers, 1933). On Curacao, a series of coarse detrital sediments overlies the limestone and is folded with it. Upper Eocene limestone is not involved in the folding. In the Dutch Leeward Islands then, a basement of volcanic rocks was deformed in pre-latest Cretaceous time, unconformably overlain by Upper Cretaceous–Eocene (?) sediments, and again folded prior to late Eocene.

Barbados–Trinidad Belt

Barbados. The geology of Barbados has been discussed in detail by Senn (1940). Clastics of early and middle Eocene age are the oldest beds exposed. These beds were uplifted, strongly folded and thrust-faulted and eroded, then covered by a thick series of mud flows. Upon the strongly folded clastic sediments and mud flows were deposited the Oceanic beds, a considerable thickness of upper Eocene chalk, radiolarian earth, and tuff, which Senn and earlier writers interpret as a deep-sea deposit. Senn shows that the area moved down very suddenly into a region of deeper-water sedimentation, a circumstance explained by great downbuckling, to be considered later in this chapter. Senn also points out that radiolarian earth similar to that of the Oceanic formation occurs in the upper Eocene of northern Cuba, and that the radiolarian earths of Barbados and Cuba probably were deposited in a late Eocene equivalent of the Puerto Rico trench. See Fig. 42.4, bottom section.

Deformation apparently continued in Barbados during the deposition of the Oceanics, for these beds are also folded and faulted, though much less so than the older formations. The Oceanic beds, in turn, were uplifted and eroded, then submerged and covered by upper Oligocene–Miocene marls. There is evidence of Miocene-Pliocene folding and post-Pleistocene uplift and fracturing. The above review was taken from Maxwell, 1948.

Tobago. The northern part of Tobago Island is made up of isoclinally folded schists, phyllites, predominantly metavolcanic in origin. South of the schists lies a belt of igneous rocks, including ultramafic and dioritic intrusives and andesitic and basaltic volcanics. A low, coral-covered plain forms the southwest tip of the island.

At least two periods of diastrophism are indicated. The earliest, probably of Late Cretaceous age, produced the schists. Intense igneous activity followed this diastrophic period; then the igneous rocks were themselves strongly sheared by diastrophic movements considered to be of late Eocene age.

Undeformed, fossiliferous upper Miocene–Pliocene sands and clays lie unconformably on volcanic rocks near the present coast line, and Quaternary coral limestone overlaps both igneous rocks and late Tertiary sediments. The above review was taken from Maxwell, 1948.

Trinidad. The middle Eocene clastic sedimentation, the late Eocene deformation, and the Miocene–Pliocene period of folding of Barbados are paralleled by a similar sequence of events in Trinidad. In addition, Juras-

sic and Cretaceous rocks crop out in Trinidad, giving insight into the pre-Eocene history of the southern West Indian region. Jurassic rocks are found only in the North Range. According to Senn (1940), they consist mainly of phyllites with abundant lenses and veins of quartz and interbedded crystalline limestone. Presumably they are equivalent to the lower Caribbean series of Waring (1926), which he describes as calcareous and carbonaceous schists and quartzitic grits. Associated with the Jurassic rocks is a younger system of less metamorphosed dark limestones, grits, and slightly metamorphosed shales, from which Trechman (1935) collected fossils of late Cretaceous age. The North Range schists have been tightly folded, in general, showing axial-plane foliation (Waring, 1926). They strike a few degrees north of east and are overturned toward the north.

In a small area near the village of San Souci, igneous rock identified as "granophyr" intrudes dark, calcareous schists of the lower Caribbean series (Waring, 1926). Both massive and pyroclastic igneous rocks are present, and on Manantial hill, dark, calcareous schist seems to be infolded into the igneous mass. A fine-grained, holocrystalline augite andesite with a diabasic texture occurs in the San Souci area.

The North Range seems to have been involved in at least two periods of folding, one in post-Jurassic and one in post-Late Cretaceous time. Igneous rocks at San Souci were intruded and extruded between the periods of deformation; otherwise there is no evidence of igneous activity. Quite probably the earlier folding, involving Jurassic and possibly Lower Cretaceous sediments, took place in Late Cretaceous time, since this is a period of major deformation in the Coast Range of Venezuela. A slight amount of volcanic activity followed this deformation in Trinidad, then uppermost Cretaceous sediments were laid down and subsequently folded, probably during the strong middle Eocene deformation.

The North Range schists of Trinidad resemble the Tobago North Coast schists in degree of metamorphism, in the fact that both series are isoclinally folded and overturned to the north, and in that both have associated younger andesitic volcanics. On the other hand, Tobago lacks the limestone, graphitic schists, and coarse grits of the North Range; and Trinidad has no counterpart of the metavolcanics comprising a major part of the Tobago schists. The small amount of igneous activity in Trinidad is likewise in marked contrast to the predominantly igneous character of Tobago. The above review was taken from Maxwell, 1948.

Northern Venezuela. Schists identical with those of the North Range of Trinidad appear in the Serrania de la Costa Oriental of Venezuela. East of Caracas, fossils of probable latest Jurassic age were found in the older beds of the Serrania de la Costa Occidental (Wolcott, 1943). Presumably the overlying schists are of Cretaceous age, as in Trinidad. The first great deformation in northern Venezuela occurred in the Cretaceous, probably in pre-late Senonian and certainly in pre-Maestrichtian time. The second deformation came about at the beginning of the upper Eocene, at which time granitic rocks were intruded in the Coast Range, and the Cretaceous and Tertiary sediments were metamorphosed (Hedberg, 1937). In the Miocene–Pliocene deformation, the Serrania del Interior was formed, involving Cretaceous and Tertiary beds in tight folding and southward overthrusting.

A belt of small serpentine intrusives occurs in the Serrania de la Costa near Caracas, and several larger peridotite bodies intrude Cretaceous sedimentary rocks south of the Serrania del Interior. Hence the ultramafics are late Mesozoic or younger in age.

PUERTO RICO TRENCH AND GRAVITY ANOMALIES

Submarine Topography

North of Hispaniola, Puerto Rico, and the Virgin Islands bank is a narrow trough of great depth. Its bottom exceeds 24,000 feet from the west end of Hispaniola eastward to a point off the island of Barbuda, a distance of about 500 miles. For a distance of 200 miles north of Puerto Rico, the trough is over 27,000 feet deep, with a greatest recorded depth of 28,680 feet. Southward from a point off Barbuda, the trough follows the arc of the volcanic Caribbees but begins to shallow, and finally it ends in the Tobago trough, a fairly wide basin betwen the volcanic arc on the west and the island of Barbados on the east. The Tobago trough has a greatest known depth of 8220 feet. Refer to map of Fig. 42.1 and cross section of Fig. 42.11.

The island of Barbados lies on a ridge that flanks the convex side of the trough and that plunges northward into deep water. Southward from Barbados, the ridge continues to Tobago, where it merges with a broad shelf off Venezuela.

Gravity Anomalies

Since Vening Meinesz's (1930) discovery of the belt of high deficiencies in gravity around the islands of the West Indies, the U.S. Navy has taken numerous gravity readings, under the direction of several scientists, and has demonstrated there a strip or belt of great negative anomalies. Its position is shown on the map of Fig. 42.9, which has been compiled by Lyons (personal communication, 1956) from all available sources. The anomaly values along the negative strip commonly reach −150 milligals, with the largest over the Puerto Rico trough north of Puerto Rico of −183 milligals. Here the axis of the negative strip is practically coincident with the axis of the trough. The negative axis extends over the Barbados ridge, however, as it is traced southwards, and over land in Trinidad and adjacent Venezuela where negative values of over −200 milligals are recorded. Another axial strip of high negative anomalies lies just north of the Dutch Leeward Islands and is about coincident with the Leeward trench (Fig. 42.1).

The anomalies are strongly positive over the Mexican, Colombian, Venezuelan, and Yucatan basins, and also over the Cayman trench (Fig. 42.9), which suggests that the Cayman trench is a different kind of tectonic feature from the Puerto Rico trench.

Concept of the Tectogene

In order to account for the belt of strong negative anomalies, generally parallel with the rises and troughs of the volcanic arcs but falling indiscriminately on one and the other, Vening Meinesz (1930) concluded that the cause was much more deep-seated than these topographic features and due to masses of lighter density material of great volume downfolded into the heavier subcrustal material. The great downbuckle is illustrated in Fig. 42.10. It was named the tectogene by Kuenen (1936). The gravity anomaly curve is also shown in Fig. 42.10; and it may be seen that the relation of the great downfold to the surficial features is direct, but that they are puny in relative size, and that the position of the negative anomaly axis to them is fortuitous. The downfold is thought by some to be driven by convection currents in the mantle, and by others the process of downfolding is thought to stimulate convection circulation. The downfold has been illustrated in model form by Kuenen (1936), and the driving mechanism and nature of surficial deformation also in model form by Griggs (1939).

In the event that the driving mechanism slows or stops, the tectogene will start to rise through isostatic adjustment, and two broad linear uplifts will appear on either side of the axis of the downfold. Pursuing this thought and mindful of the geology of the Greater and Lesser Antilles, Hess (1933) has written as follows:

> A second great deformation has occurred a considerable time after the first one, during which the tectogene originally was developed. In the interval between the first and second great deformations, one or both of the geanticlines on either side of the tectogene may have emerged above sea level. Erosion of these emergent portions, plus a great contribution of volcanics from the concave side of the arc, may deposit great thicknesses of material in "geosynclines" within the inner geanticline, and perhaps also outside of an outer geanticline, as well as in the central basin over the tectogene itself. This basin over the tectogene will henceforth be called the "geotectocline" because of its different structural behaviour and in many cases its different type of sedimentary sequence than that which occurs in a geosyncline as the term is generally used today. The second deformation will deform very intensely the material of the geotectocline. Strong folding and perhaps thrusting of the interdeformational sediments, if deposited, will occur, and probably further upthrusting of material originally squeezed out of the tectogene, if present, will take place. This happens because the material in the geotectocline is pinched between a sort of jaw-crusher as the main crust moves toward the tectogene and down over its rolling hinges. Furthermore, the material which may be on the geanticlines or in the adjacent geosynclines on the sides (or side) away from the geotectocline will be carried forward toward the geotectocline. This material may then impinge against the upsqueezed mass in the geotectocline. Upon coming against this bulwark, the weak upper part may be literally scraped off the main crust as it rides forward and down into the tectogene. This is particularly true if very incompetent horizons, such as salt beds or argillaceous sediments are contained in it. The result will be that the cover will be thrown into folds and perhaps develop a schuppen structure as its forward progress is stopped by the bulwark and the main crust under-rides or in reality underthrusts it.

Fig. 42.9. Gravity map of the Mexican–Antillean–Caribbean region, by Paul Lyons, 1956.

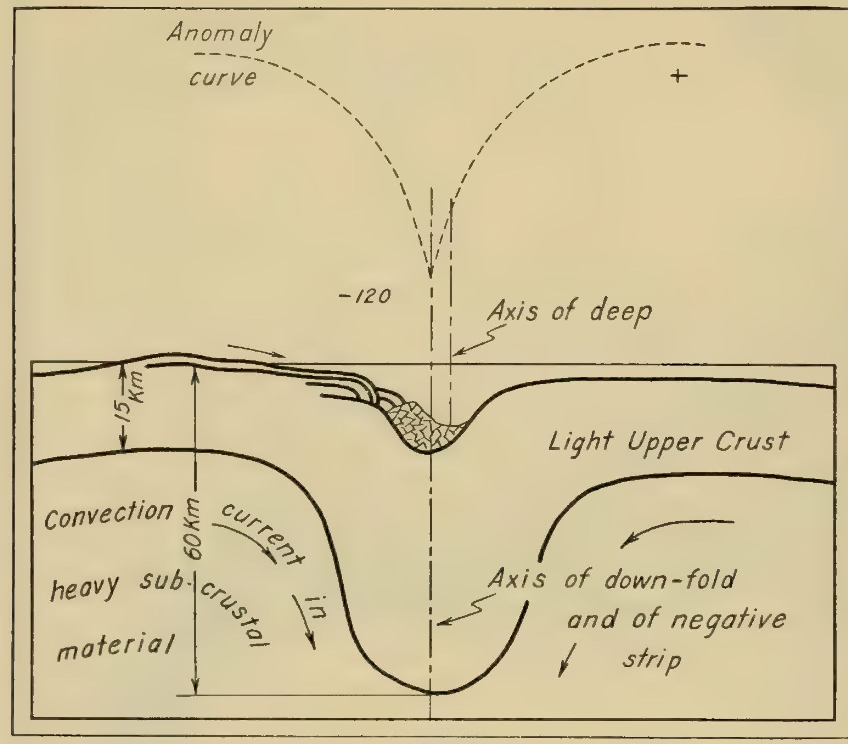


Fig. 42.10. The hypothetical tectogene and its relation to the axis of the deep and the negative anomaly strip. After Hess, 1933.

Serpentinite Intrusions and the Negative Strip

Serpentinized peridotite intrusions occur all along the great negative strip of the East Indies, and are present similarly in the West Indies. Hess (1937b) has pointed out this relation, and believes they come up along each side of the tectogene, but only in a few places are both sides exposed and not covered by younger rocks. He believes also that they are intruded only during the first great deformation producing the strip. It will be recalled that most of the serpentinized ultrabasic rocks in the West Indies previously described are part of the Late Cretaceous and Early Tertiary orogenies. Hess believes also that the serpentinites are a water-rich product of partial fusion of the peridotite substratum squeezed off as a result of the downbuckling. Conditions are favorable for its migration to the surface along the vertical limbs of the tectogene. Any later deformation does not produce such intrusion, because the bottom of the tectogene is sealed by fusion, or all the low-melting constituents of the underlying peridotitic material have been removed during the earlier cycle.

> [The serpentinized intrusions] . . . thus become useful guides in the interpretation of any region such as the West Indies. The serpentinites clearly indicate the former extension of the negative strip from the west end of Haiti along the north coast of Cuba, and thence probably to Guatemala.

CARIBBEAN REGION AND SEISMIC PROFILES

Seismic Data and General Observations

Figures 42.11 and 41.15 give the principal results of seismic exploration of the crust in the Puerto Rico region, the Lesser Antilles, the Venezuelan basin, and the Colombian basin. This significant work has been under the general direction of Maurice Ewing.

It will be noted first, that the crust of the general Caribbean region is thicker than the typical oceanic crust, but not as thick as typical continental crust; second, the Caribbean crust appears to contain no silicic basement complex characteristic of the continents, but instead, rocks interpreted mostly as volcanics. In short, the crust of this mediterranean region was once typical oceanic crust but has had unusual amounts of volcanic rocks spread widely but irregularly on it, and has suffered certain deformation. It will also be noted that under four of the basins the crust is thinner (Mexican, Yucatan, Cayman, and Venezuelan), but that under the Colombian basin, and perhaps under the Puerto Rico trench, it is thicker. The rises, ridges, and land elements are supported by thicker crust. From general isostatic considerations this is to be expected, but the negative anomaly belt and postulated tectogene under the Puerto Rico trench pose a problem.

Puerto Rico Trench

Following a widespread acceptance and intensive development of the tectogene concept, the seismic refraction survey seemed to refute the

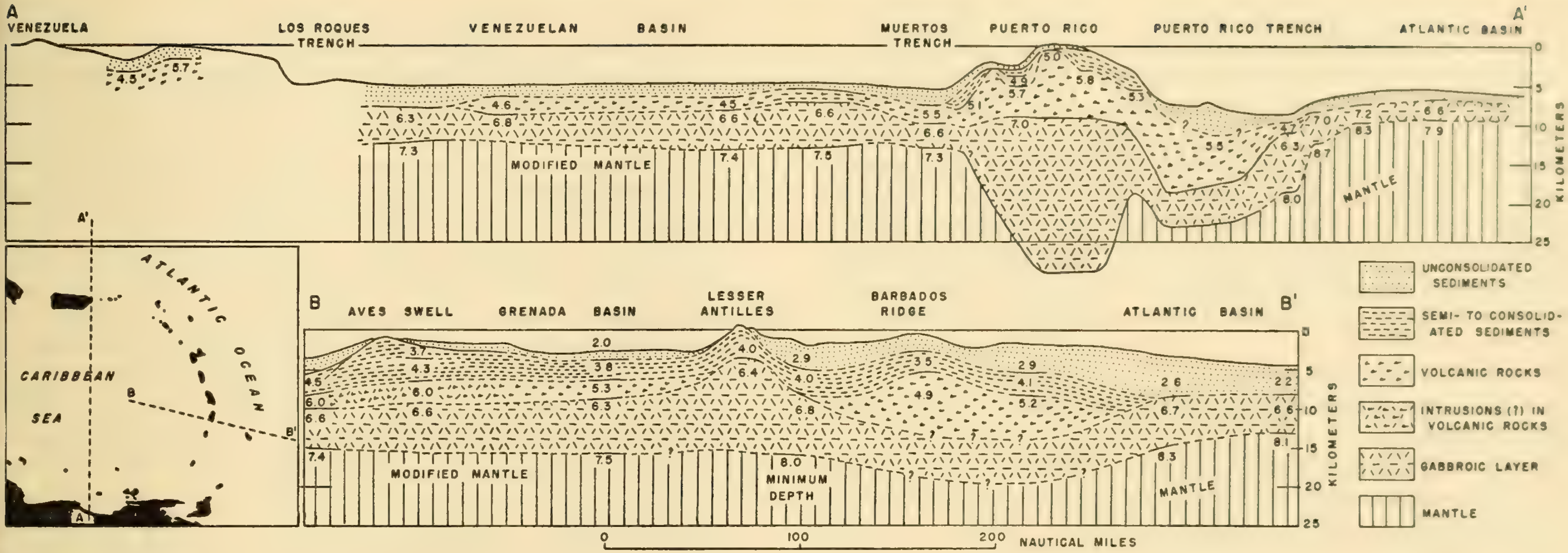


Fig. 42.11. Seismic sections and speculative geologic interpretations of the Venezuelan basin and Greater and Lesser Antilles. After Officer *et al.* (1957) and J. I. Ewing *et al.* (1957). Rocks having velocities of 4.5 km/sec., if volcanics would be porous or more silicic than basalt; those having velocities of 5.8 to 6.0 would be volcanic with intrusions of basic rock or mixtures of basalt and more silicic type, or perhaps andesite. For explanation of "modified mantle" see text. Partially modified under Puerto Rico and the Puerto Rico trench according to Talwani *et al.*, 1959.

theory in the Puerto Rico area, one of its classical localities. Ewing and Heezen (1955), Worzel and Shurbet (1955), and Shurbet and Ewing (1956) conclude from topographic profiles and sediment samples that the trench has been partly filled with several thousand feet of light, unconsolidated sediments, and that these and a thin crust can account for the belt of negative gravity anomalies equally as well as a great downfold or tectogene. The sediment fill may range up to 7 kilometers thick, depending on the thickness of the gabbroic layer assumed. They pursue the idea further by bringing to bear on the subject earthquake seismology, magnetism, and seismic refractions. No refractions had been obtained from the Moho discontinuity, or top of the mantle, under Puerto Rico or the trench, but an excellent profile had been determined across the Cayman trench (Fig. 41.15), which indicates there a marked thinning of the crust. The Cayman trench was considered by Ewing and Heezen (1955) the same kind of tectonic feature as the Puerto Rico trench, and hence, the crust should be conspicuously thin under the Puerto Rico trench. The concept of thinning under the trench is diametrically opposed to the tectogene concept. From gravity calculations the crust under Puerto Rico should be about 20 to 25 kilometers thick (Worzel and Shurbet, 1955). The gravity map of Fig. 42.9 indicates a general positive gravity area across the Cayman trench with a negative belt on the south through the northern part of the Nicaraguan rise and a mild negative belt on the north in the Bartlett deep strip. These gravity data do not agree with the computed values of Ewing and Heezen (1955), from which they deduce a conspicuous thinning of the crust. The gravity data of Fig. 42.9, are difficult, in fact, to reconcile with the refractive seismic data across the

Yucatan basin, Cayman trench, Nicaraguan rise, and Colombia basin. It appears that the refractive seismic data are to be sought, and that the gravity data are to be considered afterward in light of the seismic data.

Later refractive recordings and computations by Officer *et al.* (1957) and Talwani *et al.* (1959) in the Puerto Rico trench suggest a *thickening* of the crust there. Examine the crustal structure under the Puerto Rico trench of Fig. 42.11, section A–A′. The gabbroic and volcanic layers seem to be dropped down by a series of faults under the trench with the Moho at 20 kilometers. The gabbroic layer under Puerto Rico thickens greatly, and the Moho discontinuity reaches to a depth of about 30 kilometers there.

Lesser Antilles and Barbados Ridge

A section through the Lesser Antilles and Barbados Ridge is given in B–B′, Fig. 42.11.

Before proceeding with the interpretation of the specific velocity layers of the Lesser Antilles it will be pointed out that, in general, velocity layers are interpreted to designate rock types as follows:

1.8–3.7 km/sec	Unconsolidated and semiconsolidate sediments
4.7–5.2 km/sec	Semiconsolidated and consolidated sediments
5.2–5.5 km/sec	Extruded porous volcanic material or limestones and dolomites
5.8–6.1 km/sec	Granitic basement
6.5± km/sec	Basaltic or gabbroic subcrust
7.5± km/sec	Mixture of mantle and gabbroic subcrust or mantle with a higher temperature than general

In commenting on the velocity layers of Fig. 42.11 J. I. Ewing *et al.* (1957) conclude that a low velocity sedimentary layer consisting of an upper unit with velocity of about 1.7 km/sec and a lower unit of about 2.4 km/sec extends across most of the section. Underneath this is a layer of about 4 km/sec which also extends across all the section except under the Atlantic Basin. It could be identified as lithified sediments or porous volcanic rocks. The next higher velocity layer is one having velocities of 4.9 to 5.2 km/sec, and in interpreted to be intruded volcanic rocks. This layer is lenticular in cross section and principally under the Barbados Ridge; it does not extend under the ridge of the Lesser Antilles. However, in the Puerto Rico region a layer having velocities of 4.9 to 5.8 km/sec seems similar and has been interpreted as the Cretaceous basement of folded shales, tuffs, and agglomerates which have been extensively intruded. The layer has velocities up to 6.1 km/sec in places, and it is those parts with velocities between 5.7 and 6.1 particularly that are shown in the cross sections of this book as having considerable intrusive rock. Metamorphism incident to orogeny as well as intrusions may contribute to higher velocities in the layer, such as is evident under Puerto Rico. The most likely interpretation is that the Barbados Ridge is a large synclinal structure as far as the gabbroic layer is concerned and the low-velocity layers above the volcanic lens of the Barbados Ridge are anticlinal. The Lesser Antillean Ridge is a large anticline, or at least a belt of uplift of the gabbroic crustal layer. It appears to J. I. Ewing *et al.* (1957) that the uplift is due to extensive intrusions from below of gabbroic material. Some of this activity has penetrated through to the surface to form the present volcanoes and extrusive rock. During the course of magmatic activity differentiation to more silicic types has occurred, which are observed at the surface.

The thick lens of volcanic (?) material in the syncline under the Barbados Ridge, by the same reasoning, would be the result of an older phase of intrusive activity, but the belt has subsided incident to the new adjacent intrusive activity. The cause of subsidence is not clear. The Barbados Ridge is not active volcanically or magnetically whereas the Lesser Antillean Ridge is highly active in both respects.

Grenada Basin, Aves Swell, and Venezuelan Basin

Layers of volcanics and sediments are spread under the Grenada basin, the Aves swell and the Venezuelan basin as illustrated in section B–B′, Fig. 42.11. The gabbroic layer becomes thinner westward of the Lesser Antilles uplift and the overlying volcanics thicker. The Aves swell is apparently an especially thick pile of volcanics, and like the Barbados ridge, may have been the site of a previous uplift with intrusive activity.

Colombian Basin

The Colombian basin appears to have the same volcanic layer as the Venezuelan but the gabbroic layer is much thicker. The gabbroic layer is especially thick under the Nicaraguan rise but thins somewhat under the basin.

ORIGIN OF THE CARIBBEAN BASINS, TRENCHES, AND RISES

Magmatic Activity

It appears evident from the widespread occurrence of volcanic and intrusive rocks in the islands of the Greater and Lesser Antilles, and in the submerged areas as interpreted from the seismic layers, that the development of the Caribbean region started with oceanic crust and proceeded to evolve by abundant and widespread magmatic activity (J. I. Ewing *et al.*, 1957). Although previous theories have held the magmatic activity to be secondary to tectonic forces of compression or shear, J. I. Ewing *et al.* believe it may be the primary cause of the island arc structure. As suggested in Chapter 33 on the igneous provinces, the upper mantle is believed to melt partially at times and in places, and to yield a liquid of basaltic composition which intrudes the gabbroic layer and adds to it. In continental areas the heat released from the basaltic intrusive masses may result in the melting of lower parts of the sillicic crystalline basement and produce large volumes of monzonitic magma, but in the oceanic areas no silicic layer is present to be melted, and only fractionation of the basaltic magma can occur to produce eruptives other than basalt. Perhaps large volumes of the extruded rock is andesite. (See Chapter 33 on origin of andesite.) In the major basins perhaps the volcanics are fissure flows and mostly basalt. The variable velocities will depend on the relative amounts of basalt and andesite, on the porosity of the eruptives, and on the presence and volume of later intrusives into the volcanics.

The thickening of the gabbroic layer and the accumulation of a number of kilometers of volcanics on the thickened basalt layer, plus intrusives in the volcanics will make up a crust which stands higher than the adjacent oceanic crust. Therefore, in the manner postulated by Benioff in Chapter 32 for the orogenic belts along the Pacific margin, the higher crust of the Venezuelan and Colombian basins will tend to flow outward toward the lower Atlantic and Pacific crusts and override them. This accounts for the compressional structures at the junction and the formation of the complementary upfold and downfold (rise and trench) of the Puerto Rico and Lesser Antilles arc, and also for the arcuate map pattern of the belt of deformation (theory of J. I. Ewing *et al.* [1957] and Officer *et al.* [1957]). The overriding of thick crust on thin crust generates a shear which dips under the thick crust to great depths (Chapter 32 and Fig. 38.3) and provides an avenue for volatiles and perhaps even magma to rise further from the mantle. This engenders additional volcanic activity in the uplift inside the trench.

Since continental Venezuela and Colombia stand higher than the adjacent basins, the tendency will be for the continental margin to move northward and override the basin. The deformed belt of the Dutch Leeward Islands and the Leeward and Los Roques trenches, together with the belt of negative gravity anomalies, support this postulate. The Colombian basin crust may have tended to flow westward toward the Pacific, and the Central American trench suggests this idea (Chapter 32), but the trench continues northwestward to southern Mexico beyond the sphere of influence of the Colombian basin. Conditions are complex in Central America and will be commented on later.

Mexican, Yucatan, and Cayman Depressions

It has been pointed out that M. Ewing and associates believe the Cayman trench and the thinned crust under it to denote a structure which is the result of tension. In cross section the structure is like a necked portion of a steel rod which has been deformed under tension. The Yucatan, Mexican and southern part of the Colombian basin are structures which in line of cross section (Fig. 41.15) appear to be similar to the Cayman, but their shape in plan view must also be regarded.

The Cayman and Yucatan basins are relatively narrow and long, and are marked by strong positive gravity fields. The Mexican and Colombian basins are broad, but also are marked by positive gravity fields. The active seismic belt passes westward from Puerto Rico through Hispaniola to the

Cayman trench, along the Cayman trench and to the Gulf of Honduras, and thus it is seen that the Cayman trench of all four basins alone is seismically active. If the other basins have a similar origin, then their formation occurred in earlier times, and the Cayman should be considered in process of formation today.

The Cayman trench has long been considered a down-faulted trough, and the fault scarplike topography of the trench walls has been cited as evidence. Also, the escarpment of the Sierra Maestra of southeastern Cuba facing the trench, and the fault valley of the Cul de Sac of Hispaniola are taken to mark the eastern extent of the trench faults (Taber, 1922). The other basins have been imagined blocked out by faults (review by Eardley, 1954) but on a tenuous basis. On any grounds, the downfaulting could not affect the base of the crust which has moved up.

Ramifications of Tension Hypothesis

With the tension hypothesis before us several thoughts result: (1) How does tension in the western Caribbean relate to the outward flow and peripheral compression of the eastern Caribbean crust, the theory just proposed? (2) If these basins mark lanes of thinning of the entire crust and consequently extension, are we dealing with the drifting of South America apart from North America? (3) If the outward flow theory pertains to the eastern Caribbean, why has not the Gulf Coast of the United States flowed toward the Mexican basin and caused a volcanic archipelago and trench there? Likewise why has not the Brazilian continental margin overridden the Atlantic Ocean crust?

Continental drift and the oceanward flow hypothesis of continental margins freshly formed by fragmentation and drifting apart seem logically related, but a serious objection to the oceanward flow hypothesis as noted above, may be an argument against drifting.

POSTULATED EASTWARD SHIFT OF CARIBBEAN BLOCK

In 1938 Hess presented a theory of evolution of the Antillean region that involved eastward displacement of the Caribbean block. He regarded major horizontal shortening in the orogenic belt of the Lesser Antilles necessary as the crust was rolled down in the tectogene, and accordingly imagined the Caribbean block between Jamaica, Hispaniola, and Puerto Rico on the north and the Leeward Islands on the south to have been translated 50–100 miles eastward, and in the course of this movement the faults of the Cayman trench and the Anegada Passage were formed, and were chiefly ones of horizontal movement. Hess and Maxwell in 1953 depict some changes in the original theory (Fig. 42.12), and propose that the areas of metamorphic rock of the Greater Antilles were once joined in a single strip before the strike-slip faulting of great magnitude broke and displaced the strip. They write as follows:

> This reconstructed strip of metamorphic rocks represents the axis of the mid-Cretaceous down-buckle or downbulge. The tectonic axis is not the present negative anomaly strip north of Puerto Rico and Hispaniola as previously believed. . . . In all our previous analyses the structure of Puerto Rico appeared to be anomalous. Here the folds are overturned to the north-northeast. If the tectonic axis lay to the north, the overturning should have been southward. In our present analysis the tectonic axis lies to the south-southwest, and the Puerto Rican structures are then in a consistent relation to it.

A system of faults in northern Colombia and Venezuela are interpreted as wrench faults with movement of the Caribbean block eastward a considerable distance (Rod, 1956; Alberding, 1957). This supports the theory of Hess and Maxwell.

Bucher (1952) has presented a variation to Hess's theory. He believes that *en echelon* arrangement of fold axes along the coast of Venezuela on the south and in the islands of the Greater Antilles along the north indicates that the crust of the Caribbean Sea basin has moved eastward. Crosswise of these compressional structures is a set of high-angle faults, presumably tensional structures, which completes the picture of a shear zone along the south and north sides of the sea basin. He says:

> In the Greater Antilles, 500 miles to the north, the same combination of features recurs, but with directions reversed. There also, straight east-west trending coast lines are conspicuous in the shapes of the islands from Jamaica and the Sierra Maestra of Cuba through Hispaniola and Puerto Rico. As in the ranges that form their counterpart in the south, the axes of individual folds trend obliquely across the ranges and shore lines. But here they trend east-southeastward, while there they bear east-northeastward.
>
> A complementary set of northeast-trending fractures finds conspicuous ex-

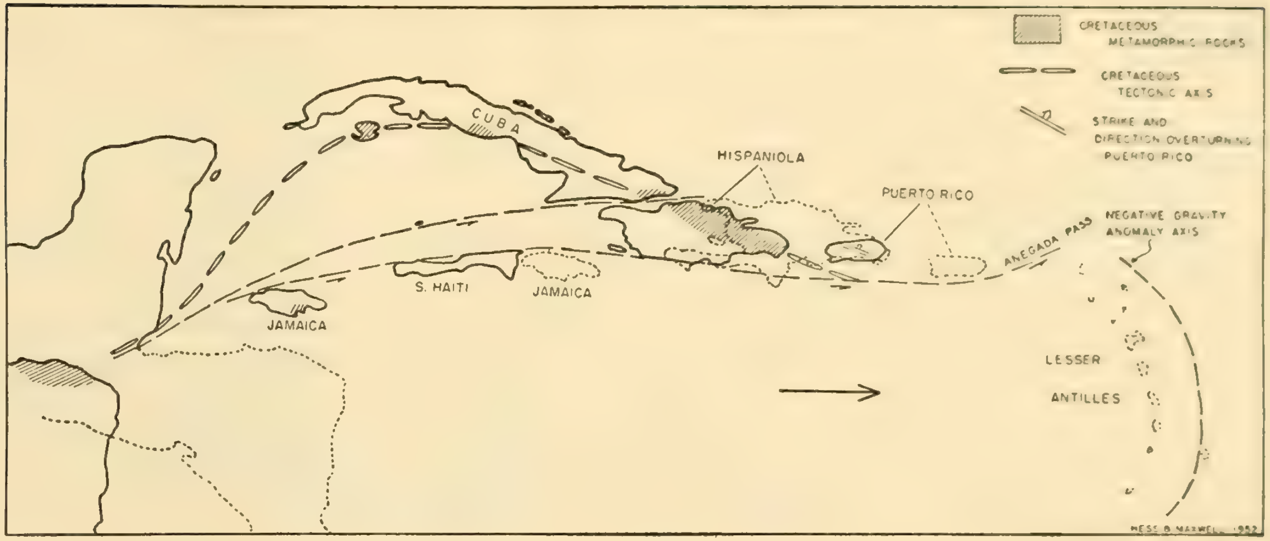


Fig. 42.12. Strike-slip faulting in Greater Antilles. Hypothesis of Hess and Maxwell (1953) showing presumed positions before and after horizontal translation.

pression in the contours of the sea floor, as in the Anegada Passage . . . and again west of St. Croix; in the Beata Ridge and its northeastward continuation along the south coast of Santo Domingo, northeast of Cape Beata; in the Navassa–Jamaica Passage (Bucher, 1952, p. 83).

Since the seismic data essentially preclude the existence of a major tectogene in the Lesser Antilles, the need for large-scale eastern movement of the Caribbean block is mostly dissipated. Also since the seismic data of the Cayman trench indicate considerable stretching of the crust, the wrench fault hypothesis hardly seems compatible with so much tensional strain. In fact, both the tectogene and wrench faults were conceived before the seismic refraction surveys, and they do not account for the crustal structure that the surveys reveal. The postulated wrench fault pattern of the Colombian and Venezuelan coast is fairly impressive, but yet some of the assumed relations are rather tenuous. In the continental drift hypothesis South America because of its present position appears to have moved eastward as well as southward while maintaining a north-south orientation. If so, considerable eastward shearing could have occurred along the Greater Antillean alignment. But at the same time the north coast of South America should have moved eastward also relative to the Caribbean block, and this is just opposite to the direction indicated by the postulated wrench fault pattern.

The origin of the Gulf of Mexico and the Antillean region is still unknown.

43.

SOUTHERN MEXICO AND CENTRAL AMERICA

MAJOR GEOLOGIC DIVISIONS

A great Cenozoic volcanic province, or possibly a complex of three or four volcanic provinces, extends through southern Mexico and Central America. The role of volcanism is most important in the geologic thinking about the region. If however, the rocks older than the Cenozoic volcanics are considered, a significant foundation geology becomes evident. A belt of crystalline rocks extensively overlain by volcanics comprises the southwestern and southern coast of Mexico, of southern Guatemala, most of Nicaragua, and all of El Salvador and Honduras. See map, Fig. 43.1. These metamorphics are assigned variously Mesozoic, Paleozoic, and Precambrian ages. A belt of deformed Permian strata with Permian (?) granitic and ultrabasic intrusives makes up part of the crystalline belt through Oaxaca, Chiapas, Guatemala, and northwestern Honduras.

A system of folds in Jurassic and Cretaceous basin-type sediments (the Mexican geosyncline in Mexico) extends along and inside the crystalline belt from southern Mexico eastward through British Honduras to the Caribbean and projects toward Cuba, the Yucatan basin, and the Cayman trench. It is called the Late Cretaceous and Early Tertiary fold belt on Fig. 43.1.

Facing the Gulf of Mexico is a narrow coastal plain which extends to the broad platform of the Yucatan Peninsula and Campeche Banks.

The crystalline and fold belts and the Coastal Plain are referred to as nuclear Central America, in contrast to the narrow volcanic province of southern Nicaragua, Costa Rica, and Panama, which is called the Isthmian link (Roberts and Irving, 1957).

CRYSTALLINE BELT

Metamorphic rocks crop out in wide areas in the Mexican State of Sinaloa which borders the southern part of the Gulf of California. Other occurrences are shown inland at Parral in the State of Chihuahua. See the new *Geologic Map of Mexico* (1956). All are labeled Mesozoic and are identified as marbles and slates. The same rocks crop out on Las Tres Marias.

Beginning at the Bahia Banderas at the west end of the Sierra Madre del Sur (maps, Figs. 35.1 and 43.1) and extending eastward through the Sierra to the Chiapas–Guatemala border is a metamorphic belt noted as Paleozoic in age on the *Geologic Map of Mexico*. The small amount of data available indicates that the rocks consist of gneisses and schists, possibly of Early Paleozoic age, and greenstone conglomerates and phyllites, possibly of Late Paleozoic age. A large batholith in eastern Oaxaca and Chiapas is intrusive into the metamorphic rocks and is considered as Paleozoic in age by Roberts and Irving (1957) and as Mid-Paleozoic by de Cserna (1958 and 1960). The metamorphics are also intruded by Mid-Cretaceous (Cenomanian) stocks and batholiths which are probably

Fig. 43.1. Tectonic map of southern Mexico and Central America. Compiled from Roberts and Irving (1957), de Cserna (1958), Terry (1956), and *Geological Map of Mexico*, I. G. C. (1956). For distribution of Tertiary volcanic rocks in southern Mexico see Fig. 32.8. Southern Guatemala, El Salvador, southern Honduras, central and southern Nicaragua, Costa Rica, and Panama are nearly all covered by Cenozoic volcanic rocks. B.H., British Honduras.

related to the Nevadan orogenic belt (see Chapter 35). The older intrusions are overlain in Chiapas by late Paleozoic sediments and locally elsewhere in the Sierra Madre del Sur by Lower Cretaceous and Jurassic marine strata. A few scattered observations of the direction of foliation seem random, and so no conclusions are yet justified regarding the trends in the metamorphic complex (de Cserna, 1958).

In northern Honduras and Nicaragua the metamorphic belt is labeled "probably Precambrian" on the *Geologic Map of Central America* (Roberts and Irving, 1957), and it consists of undifferentiated schist, gneiss, phyllite, quartzite, and marble. Sapper (1937) has emphasized that the major structures here are pre-Permian.

PERMIAN FOLD BELT

The *Geologic Map of Central America* shows a belt of folded Permian strata across central Guatemala. The Chochal limestone and Santa Rosa limestone, conglomerate, shale, and sandstone are the formations identified. They are intruded by granite and serpentine, which may be Late Permian or Triassic in age, and were involved in the orogeny in which the Permian rocks were folded. The rocks of the Santa Rosa formation become progressively more metamorphosed to the east so that shale becomes phyllite and schist in eastern Guatemala and in Honduras. The Jurassic Todos Santos formation rests on the folded Permian rocks.

The crystalline complex was widely blanketed by Upper and Lower (?) Cretaceous rocks in Honduras and Nicaragua, and these have been folded in the Late Cretaceous and Early Tertiary orogeny.

An area of outcrop of the Santa Rosa formation with granitic intrusions occurs north of the main Permian fold belt in British Honduras. This may indicate that the original Permian fold belt was once wider than now, and that the later Cretaceous fold belt largely covers it.

LATE CRETACEOUS AND EARLY TERTIARY FOLD BELT

General Characteristics of Mexican Fold Belt

The strata of the Mexican geosyncline (Chapter 28) extend into southern Mexico and overlap broadly southward on the crystalline belt. These sedimentary rocks are chiefly Jurassic and Lower Cretaceous carbonates, with some fine clastics near the base, and Upper Cretaceous shales. The Albian and Cenomanian seas advanced widely over the southern and southwestern crystallines and deformed Paleozoic rocks which had previously been land and the source areas for the Jurassic and earlies Cretaceous sediments (de Cserna, 1958). At places, however, Lower Jurassic strata rest on the crystallines, as in Oaxaca. Laramide compression then deformed the Jurassic and Cretaceous strata and a long system of folds resulted. These are labeled on Fig. 43.1 the Late Cretaceous and Early Tertiary fold belt. For the most part the folds are asymmetrical toward the northeast. The fold belt is broad in northern and central Mexico (Chapter 28), but narrows southward and is marked essentially in southern Mexico by the Sierra Madre Oriental (de Cserna, 1958).

Red Conglomerates in Central and Southern Mexico

A number of occurrences of Early Tertiary red conglomerates have been noted in central and southern Mexico (Edwards, 1955). They are particularly important in deciphering the Laramide and Tertiary history of the fold belt. The localities where the red conglomerates are known are shown on the map of Fig. 43.2. The three areas studied are noted on Fig. 43.1 where they may be seen in relation to the crystalline belt and fold belt. They are the Zacatecas, the Guanajuato, and the Taxco.

The oldest rocks in the Guanajuato City area are folded, hard, black, thin-bedded, marine shales which now appear in places as phyllites or schists. Small quantities of limestone, sandstone, and volcanics appear in the series. No fossils have been found but on the basis of lithologic similarity they have been correlated with the Upper Triassic shales at Zacatecas City (Edwards, 1955). Six miles northwest of Guanajuato the La Luz schist occurs which contains an amygdaloidal basalt about 1000 feet thick. A dense, dark gray limestone, possibly of early Cretaceous age, is believed to have once covered the Triassic (?) shales, but was removed locally before the overlying conglomerates were deposited.

After the full development of the Mexican geosyncline the main Laramide orogeny occurred. Folds are the main exhibit (see Chapter 28) but here in south-central Mexico, considerable plutonism occurred. A deeply

Fig. 43.2. Map of Mexico showing localities in which red conglomerates are known to occur. Reproduced from Edwards, 1955.

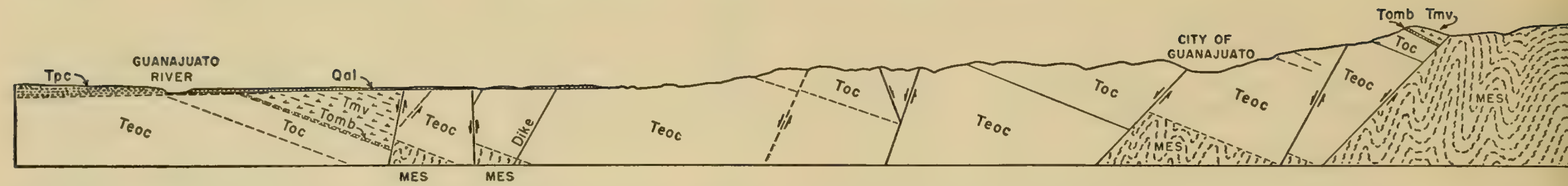


Fig. 43.3. Cross section at city of Guanajuato, Mexico, after Edwards, 1955. Qal, Alluvium; Tpc, Pliocene conglomerate; Tmv, Young volcanic rocks, Miocene; Tomb, Bufa ss., Oligocene-Miocene; Toc, upper part of Guanajuato conglomerate; Teoc, lower part of Guanajuato conglomerate; upper Eocene-Lower Oligocene; MES, Mesozoic sedimentary rocks and coarse-grained silicic, intrusive rocks.

weathered granite is intrusive into the Triassic (?) shales and schists. Other plutons are dioritic and monzonitic in composition, and all are believed to be pre-conglomerate by Edwards.

Uplift, faulting, erosion, and volcanism followed the folding and intrusions. Although the volcanics do not occur interstratified in the sequence a great pile of them is believed to have existed nearby because derived fragments constitute a large part of the conglomerate.

The red conglomerate at Guanajuato City is named after the city. There it is about 5000 feet thick but thins northeastward and southwestward. Volcanic fragments form more than half of the deposit, with granite, diorite, limestone, and chert making up the rest. Granite fragments increase upward and compose 35 percent of the mass near the top. This indicates an increasing exposure of the granitic pluton in the source area.

The source of the conglomerate was a highland northeast of Guanajuato City where silicic volcanics capped shales and limestones of Cretaceous age. The highland was also an area where a granitic pluton had intruded the Cretaceous strata (Edwards, 1955).

The Guanajuato conglomerate is late Eocene or early Oligocene in age. Similar red conglomerates presumably of the same age occur at Zacatecas and Taxco.

The conglomerate is overlain by the tuffaceous Bufa sandstone into which it is transitional. The sandstone is about 50 feet thick.

The great Miocene (?) volcanic epoch followed, which was initiated by the deposition of massive, bedded tuffs more than 1500 feet thick in the Guanajuato area. Then followed normal faulting, which produced the tilted block and graben structure so strongly evident today. The slip on several of the northwest striking faults is as much as 3000 feet. See cross section of Fig. 43.3.

Fold Belt in Central America

The fold belt is shown to include gentle folds in the states of Tabasco and Vera Cruz along the north side in a region recognized as coastal plain by some.

As the fold belt is traced eastward into Guatemala and British Honduras, continental Jurassic and Lower Cretaceous beds are involved. These beds probably covered much of the "nuclear region" of Central America (Imlay, 1944; Roberts and Irving, 1957). Toward the close of Early Cretaceous time subsidence and marine embayments resulted in the deposition of limestone and dolomite approximately coextensive with the underlying terrestrial beds. Deposition continued in most places until Late Cretaceous when the folding occurred.

The earliest Tertiary beds are coarse clastic rocks of the Sepur formation, whose composition shows that they were derived from a wide variety of sources including crystalline rocks, Permian limestone and quartzite, and limestone and volcanic rocks of Mesozoic age. The Sepur strata were probably deposited during orogenic movements in Late Cretaceous and Eocene time. Intrusions of granodiorite and diorite that accompanied the orogeny cut the Cretaceous rocks throughout eastern Guatemala and Honduras. Folds, largely trending eastward, were developed in the Cretaceous rocks and appear also to have involved rocks as young as those of Sepur age.

The orogenic movements culminated in thrust faulting, first mapped in the Departamento de Huehuetenango, which thrust the Permian rocks over the Todos Santos formation and the Cretaceous limestone. The extent of the thrusting is not known, and many such faults may be present in central Guatemala (Roberts and Irving, 1957).

Two distinct phases of orogeny are represented. First, the deformation that resulted in the deposition of the Sepur strata, and second the involvement of the Sepur in folding and thrusting as noted in Guatemala in the Departamentos de Huehuetenango, Alta, Verapaz, and Peten. The last phase must therefore be as late as Eocene or possibly post-Eocene. Miocene strata overlap the earlier Tertiary rocks on the Peten area and are hardly deformed.

The *Geologic Map of Central America* (Roberts and Irving, 1957) shows considerable Cretaceous strata lying on the crystallines of Honduras and northern Nicaragua, and also that the Cretaceous strata have been folded. The obvious fold axes are indicated on Fig. 43.1. It is therefore evident that the Laramide fold belt spread southward in this part of Central America and involved the crystalline belt somewhat.

SOUTHERN GULF COASTAL PLAIN

The Southern Gulf Coastal Plain is made up of marine Cenozoic sediments and some volcanics, and it extends as shown on Fig. 43.1 along the east side of the Sierra Madre Oriental southward through the State of Vera Cruz and thence in a narrow belt eastward through Tabasco to the Yucatan Peninsula. It is divided into a number of basins partly for the convenience of petroleum exploration, and the boundary of these basins with the fold belt is not well defined nor usually agreed upon (Benavides, 1956; Guzmán, 1959). See Fig. 41.9.

The strata of the Coastal Plain dip gently toward the Gulf of Mexico. In the Coatzacoalcos region, or the Isthmus (of Tehuantepec) saline basin salt intrusion structures are prominent, and to the east, gentle folding is prevalent.

YUCATAN PENINSULA

The terrane of the states of Tabasco, Campeche, and Yucatan is underlain by flat or very gently folded marine strata that range in age from late Eocene to Pleistocene. On the whole, the Yucatan Peninsula consists of lowlands under 650 feet in height. As the strata are mostly limestones, the country is almost destitute of rivers, and the rains sink quickly through the soluble and karsted limestone and gather in subterranean basins. The water table lies at various depths down to 300 feet or more beneath the land surface.

Yucatan extends north beneath the Gulf of Mexico for at least 150 miles as the Campeche or Yucatan Bank, and then descends abruptly into the depths of the Gulf. Carbonaceous sediments are accumulating on the shelf (Fig. 43.4).

VOLCANIC FIELDS AND FAULTING

Southern Mexico and Central America are particularly noted for volcanoes, and volcanic rocks cover extensive areas. The map, Fig. 33.6, shows the volcanic rocks of southern Mexico, where two general ages are recognized, the mid-Cenozoic and the late Cenozoic. The older volcanics make up the southern end of the extensive Sierra Madre Occidental province, but the younger eruptives are in the form of an east-west belt of stratovolcanoes including such well-known cones as Arizaba, Popocatepetl, Ixtaccihuatl, Paricutin, and Colima. The belt of stratovolcanoes has been called the Trans-Mexican volcanic belt (de Cserna, 1958), and from the map of Fig. 43.1 it can be seen to extend from the Bahia Banderas to the Gulf of Mexico below the city of Vera Cruz.

Beginning in Chiapas, not far from the east end of the Trans-Mexico volcanic belt is another great belt of modern volcanoes which stretches through southern Guatemala, El Salvador, southern Nicaragua, and western and central Costa Rica. In El Salvador major faults trend northwestward parallel to the volcanic chain and to the coast line. Lago de Ilopango occupies a graben that developed over a long period of time and was partly filled by a succession of lavas and pyroclastics during its formation. Other lake basins such as Lago de Atitlan and Lago de Amatitlan in Guatemala also probably formed, at least partly, by collapse (Roberts and Irving, 1957). The extrusive rocks are olivine basalt, basalt, labradorite andesite, and in lesser distribution dacite (Weyl, 1956). A transverse zone of faults across Honduras from the Gulf of Fonesca to the

Fig. 43.4. Sediments of the shelves around the Gulf of Mexico. Reproduced from Atwater and Forman, 1959. Bahama Banks are also calcareous.

Caribbean is shown on the *Map of Mexico,* and the Valle de Comayagua is regarded as formed by block faulting. Parallel structures along the coast of British Honduras may be related to this transverse fault zone. Uplift accompanied the volcanism over much of Central America (Roberts and Irving, 1957).

Reference has been made to the numerous submarine volcanic cones on the Pacific floor adjacent to Central America, and also to the Central American trench, in Chapter 32. See especially Fig. 32.5.

ISTHMIAN VOLCANIC LINK

Southernmost Nicaragua, Costa Rica, and Panama constitute the so-called Isthmian link. By it the more broad and massive foundations of North and Central America are connected to South America. The rocks of the link are largely igneous and stratified deposits derived mostly from igneous formations.

The Cordillera de Talmanca in central and southeastern Costa Rica has been studied in considerable detail by Weyl (1956, 1957) and the table of Fig. 43.5 gives the history of the isthmus there as he depicts it. The Cordillera Central referred to in the table is just west of the Cordillera Talamanca and is the eastern end of the stratovolcanic belt.

The record in the Sierra Talamanca goes back only to the Oligocene, but as may be seen, it is one of sinking and volcanism; folding, erosion, and deposition; intrusive activity with much metamorphism; more folding; and finally uplift, faulting and more volcanism and the building of the modern cones.

The geology of Panama begins with a basement complex of Eocene and possibly pre-Eocene (Terry, 1956) or Cretaceous (?) age (Woodring, 1957) which crops out over half the country. It is made up predominantly of altered basic flows, agglomerates, tuffs, and diorite intrusions. It is strongly deformed but little metamorphosed, although argillites are known and schist float has been reported in two places. The areas of outcrop of the basement complex are anticlines or horsts, and their axes are shown on Fig. 43.1.

EPOCH		DEPOSITS	STRUCTURAL EVENTS	MAGMATISM
PLEISTOCENE		River gravel and sand	Uplift, faulting, arching	Later volcanism in the Cordillera Central
PLIOCENE		Suretka conglomerate	Relative quiet	
MIOCENE	U.	Molasse sediments on mountain flank	Folding of border zone	Late orogenic intrusive activity
	M.	Gatun conglomerate	Main folding in the mountains	
	L.	Limestone, marl	Geosynclinal sinking	
OLIGOCENE		Tuff Chert Sandstone Marl Tuff Limestone		Beginning of basic volcanism
EOCENE		Unknown foundations		

Fig. 43.5. History of the Cordillera de Talamanca in Costa Rica. After Weyl, 1956.

The strong deformation of the Cretaceous lavas and sediments occurred in Late Cretaceous, Paleocene, or early Eocene time (Woodring, 1957).

The basement complex is overlain on the flanks of the uplifts by sediments and volcanics of several kinds, carrying fossils in many places. Fossil collections indicating late Eocene, Oligocene, early, mid-, and late Miocene, Pliocene, Pleistocene, and Recent have been described (Terry, 1956; Woodring, 1957). A section across the isthmus at the Canal Zone is shown in Fig. 43.6.

Volcanism again reached a climax during Oligocene and early Miocene time. The rocks have been identified as diorite, quartz diorite, dacite, andesite, and basalt (Woodring, 1957).

A number of trans-isthmian faults have been postulated by Terry (1956). These are shown on Fig. 43.1. Two of them in the Canal Zone and one farther west have strike slip movement, and two in eastern Panama are high-angle thrusts. The Panama Canal Zone seems to be the most complicated area, but this may be the result of more intensive field work there than elsewhere. In general it is more de-

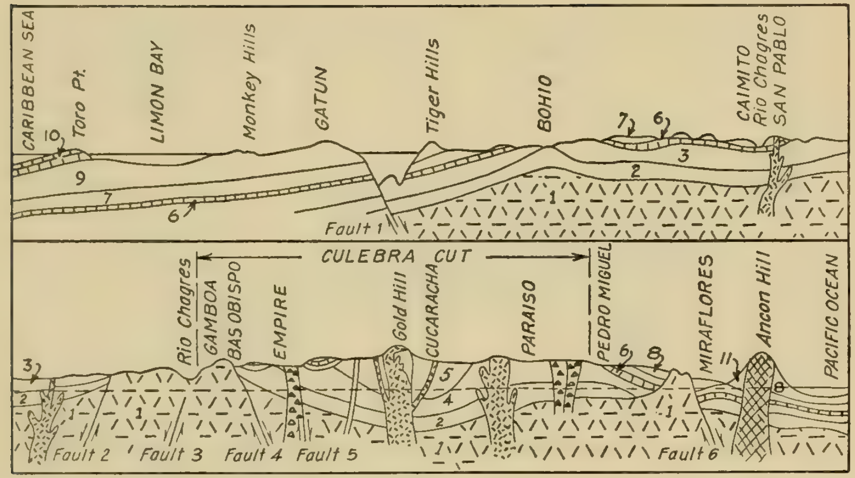


Fig. 43.6. Generalized section across Isthmus of Panama, after MacDonald, 1919. Intrusions are granodiorite, diorite, andesite, rhyolite, and basalt. Bedded rocks: 1, Bas Obispo volcanic breccia; 2, Las Cascades agglomerate (Oligocene); 3, Bohio conglomerate; 4, Culebra fm.; 5, Cucuracha fm.; 6, Emperador ls.; 7, Caimito fm. (Miocene); 8, Panama fm.; 9, Gatun fm. (Pliocene); 10, Toro ls.; 11, Pleistocene.

pressed than on either side with the basement complex showing to a less extent. A somewhat different fault interpretation of the Canal Zone has been rendered by Woodring (1957).

The Isthmian link has the shape of an S curve and the faulted area of the Canal Zone is in the middle, which may indicate that deformation has been in one direction on one side and opposite on the other (Terry, 1956).

The basement complex of Cretaceous (?) age with dioritic intrusions has been taken to continue the Nevadan orogenic belt into South America (Eardley, 1954). The broad nature of the platform upon which the exposed narrow isthmus rests has been regarded as wide enough to contain a major belt of deformation. Evidence to the contrary may be cited as follows. The intrusions are probably not batholithic in size and some may be Oligocene or Miocene in age. Also it is evident now from seismic refraction work that modified oceanic crust lies on the Caribbean side of the link, and true oceanic crust on the Pacific, so if the isthmus represents an orogenic belt, it has evolved from oceanic crust, and should not be similar to the Nevadan which in general evolved from a great eugeosynclinal complex along the margin of the continents of North and South America.

RELATION TO GREATER ANTILLES

Projection of Crystalline and Fold Belts

The Crystalline Belt through the coastal ranges of northern Honduras and Nicaragua passes out into the Caribbean Sea, and it is inferred from exposures of the crystalline rocks on the Isla Roatan that the belt continues toward Jamaica, bounding the Cayman trough on the south (Roberts and Irving, 1957). The seismic traverse of Fig. 41.15, reveals a layer of 5.2 to 6.1 kilometers per second, and although this has been interpreted as a volcanic layer, the parts with higher velocities could be the rocks of the Crystalline Belt. It seems hardly thick enough, however, to represent an old orogenic belt of metamorphic rocks. Metamorphic rocks of the Crystalline Belt varieties are not exposed on Jamaica, but rather the oldest core rocks are folded volcanics of Late Cretaceous age. It is tentatively concluded, therefore, that the Crystalline Belt as a crustal layer, wedges out not far east of the east coast of Nicaragua.

The fold belt has commonly been projected north of the Yucatan Basin to Cuba, and the Cockscomb Mountains in British Honduras project eastward toward the Misteriosa and Cayman banks (Roberts and Irving, 1957). If connections north or south of the Yucatan Basin ever existed with Cuba, they must have been very transitory, according to vertebrate paleontologists (Schuchert, 1935). Yucatan has about seventy species of verterbrates which are of the fauna of the Atlantic neotropical realm. If it was united with Cuba at any time during the late Cenozoic, it is inconceivable why tortoises, pit vipers, Opisthoglypha, and Cnemidophorus should not have crossed over in to Cuba. Since Yucatan was beneath the sea during most of the Cenozoic era, land connections through the peninsula, at least, seem impossible.

Meaning of Yucatan and Cayman Depressions

The thin crust under the Yucatan and Cayman troughs, as previously explained, might indicate immediately to the proponent of continental drift, that the crust has been stretched and is in the process of being fragmented or pulled apart. Even one not ready to accept large scale drift of the continental masses may concede some pulling apart and thinning. If stretching and thinning is admitted as a possibility, then the thin gabbroic layer under the Mexican basin may also represent drifting apart there, but at a somewhat earlier time because of the thicker layers above of consolidated and unconsolidated sediments. Right off, however, the seismic evidence does not suggest that we are dealing with a silicic layer in the Cayman region—simply the gabbrioc layer is being thinned. In continental drift, if it occurs, are we dealing with the movement of the total crust over the mantle with a new gabbroic layer forming immediately in the breach, and progressively as it widens, or does the silicic layer slide over the gabbroic?

Postulated Wrench Faults

With the discovery of the great fracture zones in the eastern Pacific the attempt has been made to connect one of them, the Clarion, with the Cayman trench across Central America. Hess and Maxwell's (1953) postulate of Fig. 42.12 shows Honduras, Nicaragua, and El Salvador to have moved eastward over 200 miles from a former position in what is now the Pacific Ocean, along a break that would transect Central America from southern British Honduras to the Pacific coast at about the Chiapas–Guatemala border. Such a break would transect the Permian fold belt and the metamorphic belt, for which there is no existing geologic evidence. Also the Central American trench is a late Cenozoic feature which, because it is continuous, precludes horizontal translation of crustal blocks across it during this time. Since the Cayman trench is a late Cenozoic feature, the movement postulated by Hess and Maxwell has to be late Cenozoic, which is impossible across Central America. The Clarion fracture zone takes off northwest of Acapulco and does not line up with a projection of the Cayman trench.

The zone of stratovolcanoes across southern Mexico might also be regarded as the line of horizontal movement, but no geologic relations are known there to denote a wrench fault zone other than the aligned vents.

MAMMALIAN FOSSIL RECORD AND LAND CONNECTIONS

The record of mammals, both existent and fossil, in North and South America and in the isthmus itself is impressive, and speaks more positively of land connections than the physical, but still the two lines of evidence lead to parallel and supporting conclusions.

South America may have had southern and eastern connections during the Mesozoic, but since late Cretaceous time at least, it has been isolated from all the rest of the world, except for occasional connections with North America (G. W. Simpson, personal communication). This has established it as an ideal laboratory in experimental evolution over a lapse of seventy million years, and the record is remarkably clear. By reference to the chart of Fig. 43.7, it will be seen that a group of early immigrants were left isolated and proceeded to evolve in their own way. The connection between North and South America to permit the influx of these early mammals evidently was the result of volcanism and the disturbances just reviewed.

In late Eocene and early Oligocene time, shallow seas and volcanic islands in the area of the isthmian uplift allowed certain forms adequately equipped to make passage from island to island, and thus a wave of "island hoppers" entered South America. After certain adjustments with the ancient immigrants already there, the newcomers also proceeded to evolve their own way. No continuous land of any breadth or durability was established at this time, and the migration route served as a screen or sieve to a host of North American forms which would have moved in under more suitable environmental conditions.

Orogeny and extensive volcanism again convulsed the isthmian swell in late Tertiary and Quaternary time, and at first the region seems to have been a chain of islands permitting a second wave of "late island hoppers," and then a solid subaerial connection, permitting a wave of new immi-

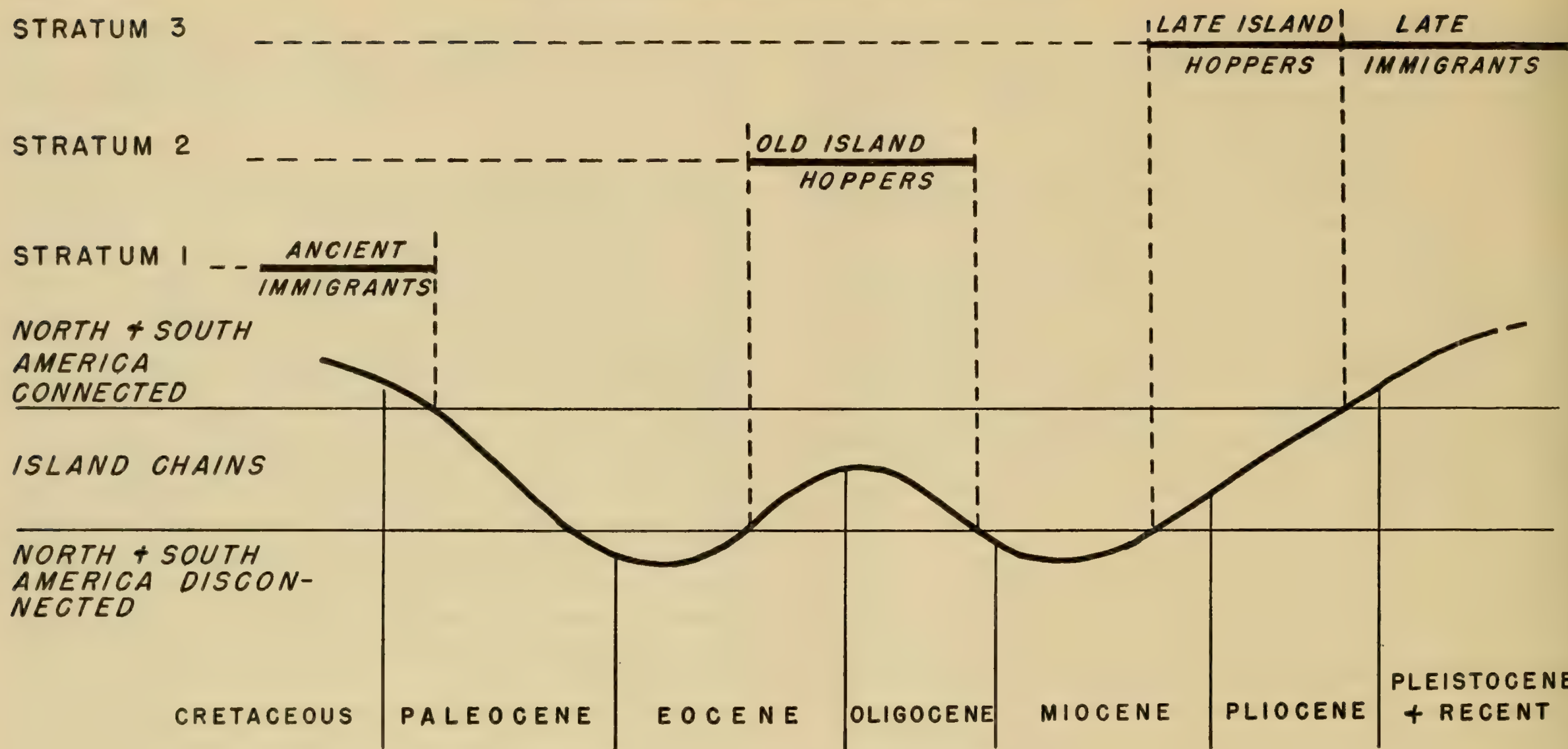


Fig. 43.7. Times of migrations between North and South America. Copied from lantern slide of G. W. Simpson, 1950 Sigma Xi lecture.

grants from North America. Some of the peculiar forms from South America also made their way into North America.

It is possible that the Antillean arc from latest Cretaceous time on could have been a land connection as well as the Costa Rica–Panama isthmus, and that their histories may not have run exactly parallel. With the possibility of two bridges between continents, ocean-to-ocean migration may have been delayed for a while in the mediterranean between bridges, in the manner of a ship negotiating locks in a canal. Also, if the history of the two bridges did not run parallel, then opportunities for exchange of land animals would be more frequent than if only one bridge had recurring emergences and submergences. The paleontologic record of the Greater and Lesser Antilles, however, does not indicate that the eastern orogenic belt was of importance at any time as a land bridge between the continents.

Permian reptiles and flora were isolated and did not migrate from one continent to the other (Schuchert, 1935). The separation continued through the Triassic. Little can be said of the Jurassic and Early Cretaceous.

At the same time as land migration routes are established between North and South America, so are migration routes of marine invertebrates severed between the Atlantic and Pacific. The conclusions reached by the invertebrate paleontologist should therefore dovetail those of the vertebrate paleontologist. According to Schuchert (1935) not all invertebrate paleontologists agree on the relation of Atlantic or Gulf and Pacific forms, but most evidence points to a portal in Early and Middle Tertiary time, and thus supports the mammalian record.

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